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Studies on Waves of Finite Amplitude*

By

M. TOMINAGA**

Abstract: Surface waves of finite amplitude generated in a canal of constant depth have been treated by Struik in 1926, but the author tried to solve the same problem by means of the Dirichret's method which led to a non-linear integral equation. Wave velocity obtained from Struik's formula is considered to be small from the physical point of view but the author's formula is valid in this respect. Wave profiles were also studied theoretically.

§1. Fundamental Equations and Boundary Conditions.

We consider a two dimensional irrotational surface progressive wave motion, whose amplitude is not infinitesimal, in a canal of constant depth. Notations employed through this paper are follows:

- λ: wave length
- φ: velocity potential
- ψ: stream function
- H: mean depth of the canal
- H': wave height (above the mean water level)
- c: wave velocity

The motion is confined to two dimensions, i.e. xy-planes. The axis of x is taken horizontal, that of y vertical and positive downwards, the origin taking on a crest of waves. The xy-plane is transformed into the $\varphi\psi$-plane as shown in Fig. 2, the free surface corresponding to the axis of φ, and the bottom corresponding to $\psi=cH$. The domain which lies between two infinite lines, φ-axis and $\psi=cH$ is represented in an annular domain D in ζ plane, through $\zeta=e^{2\pi i/\lambda c}(\varphi+i\psi)$ as shown in Fig. 3. Then the free surface corresponds to an unit-circle and the bottom to a circle of radius $q=e^{-2\pi\psi/\lambda c}=e^{-2\pi H/\lambda}$.

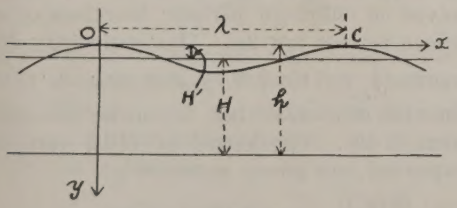

Fig. 1. xy plane.
H mean water depth.
λ wave length.

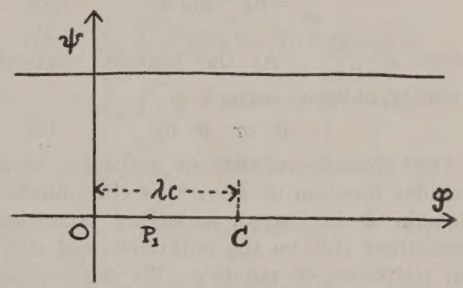

Fig. 2. $\varphi\psi$ plane.
O, C wave crest.

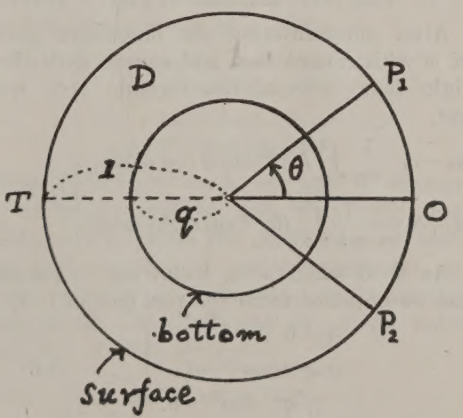

Fig. 3. $\zeta=\rho e^{i\theta}$ plane.
O wave crest.

* Received Aug. 20, 1955.

** Tokyo University of Liberal Arts

The function ω is defined[1] by

$$ce^{-i\omega}=\frac{d}{dz}(\varphi+i\psi)\ ;\quad z=x+yi$$

in the domain D, ϑ being its real part and τ being imaginary part, and ω is defined as a complex velocity; then

$$w=\frac{d}{dz}(\varphi+i\psi)=u-iv\ ;$$
$$u=ce^{\tau}\cos\vartheta,\quad v=ce^{\tau}\sin\vartheta \qquad (1\cdot1)$$

ϑ and τ are functions of θ, and because of the symmetrical character of waves absolute value of u and v must be equal for θ and $-\theta$, then from (1·1) $\vartheta(\theta)$ must be an odd, and τ an even function.

On the free surface we have $\frac{1}{2}w^2-gy=$ const. and according to Levi-Civita, this formula leads to[2]

$$\frac{d\tau}{d\theta}=Ke^{-3\tau}\sin\vartheta \qquad (1\cdot2)$$

here, $K=\frac{\lambda g}{2\pi c^2}$. At the bottom, vertical velocity must be zero;

$$v=0\quad\text{or}\quad\vartheta=0. \qquad (1\cdot3)$$

Our problem consists in getting ω as a regular function of $\zeta=\rho e^{i\theta}$ in the annular domain D in Fig. 3, satisfying boundary conditions (1·2) on the unit circle, and (1·3) on the circle of radius q. We can expand ω in Laurent's series in D:

$$\omega=a_0+\sum_{n=1}^{\infty}a_n\zeta^n+\sum_{n=1}^{\infty}b_n\zeta^{-n},$$
$$a_n=\alpha_n+i\beta_n,\quad b_n=\gamma_n+i\delta_n. \qquad (1\cdot4)$$

After differentiating the imaginary part of ω with regard to θ and equate with the right hand side of the formula (1·2) we get,

$$\left.\begin{aligned}\alpha_n-\gamma_n&=\frac{1}{n\pi}\int_0^{2\pi}Ke^{-3\tau}\sin\vartheta\cos n\theta\,d\theta\\ \beta_n+\delta_n&=-\frac{1}{n\pi}\int_0^{2\pi}Ke^{-3\tau}\sin\vartheta\sin n\theta\,d\theta.\end{aligned}\right\}\qquad(1\cdot5)$$

As $\vartheta=0$ when $\rho=q$, following formulae can be obtained from the real part of (1·4);

$$\left.\begin{aligned}&\alpha_0=0\\ &\alpha_nq^n+\gamma_nq^{-n}=0\\ &-\beta_nq^n+\delta_nq^{-n}=0.\end{aligned}\right\}\qquad(1\cdot6)$$

Solving α_n, β_n, γ_n, δ_n from (1·5), (1·6) and putting into (1·4) we get

$$\omega=i\beta_0+\frac{K}{\pi}\int_0^{2\pi}Ue^{-3\tau}\sin\vartheta\,d\theta \qquad (1\cdot7)$$

where $$U=\sum\frac{\zeta^ne^{-in\theta}-q^{2n}\zeta^{-n}e^{in\theta}}{n(1+q^{2n})}. \qquad (1\cdot8)$$

This is a solution of Dirichret's problem and called a Villat's formula, β_0 being an arbitrary constant which will be decided in later. If we introduce a Weierstrass' elliptic function $\vartheta_1(s)$, U will be represented as follows.

$$U=\log\frac{iA_q}{t}\frac{\vartheta_1(q,s)}{[\vartheta_1(q^2,s)]^2},$$
$$A_q=\frac{q^{1/6}[\vartheta_1'(0,q^2)]^{2/3}}{[2\pi\vartheta_1'(0,q)]^{1/3}}. \qquad (1\cdot9)$$

Here,

$$\vartheta_1(q,s)=2q^{1/4}(\sin\pi s-q^2\sin 3\pi s+q^6\sin s\pi s-\cdots\cdots)$$

or $$\vartheta_1(q,s)=-iq^{1/4}(t-t^{-1})\prod_{n=1}^{\infty}(1-q^{2n})\times\prod_{n=1}^{\infty}(1-q^{2n}t^2)(1-q^{2n}t^{-2})$$

and $$t^2=\zeta e^{-i\theta},\quad t=e^{s\pi i},\quad s\pi=\frac{\sigma-\theta}{2}.$$

In our case a point ζ is considered to be fixed, so σ is a fixed angle and θ is a current angle. Taking the real part of (1·7) we get,

$$\vartheta(\sigma)=\frac{K}{\pi}\int_0^{2\pi}e^{-3\tau}\sin\vartheta\cdot\log\left|A_q\frac{\vartheta_1(q,s)}{\vartheta_1^2(q^2,s)}\right|d\theta. \qquad (1\cdot10)$$

From (1·2) we get

$$\tau=\int_0^{\sigma}Ke^{-3\tau}\sin\vartheta\,d\theta \qquad (1\cdot11)$$

(1·10) and (1·11) belong to a system of non-linear integral equations, τ and ϑ are solved as unknown periodic functions of σ whose periods are 2π. This periodicity requires $\int_0^{2\pi}e^{-3\tau}\sin\vartheta\,d\theta=0$ and also A_q is a function of q only, then A_q can be removed from (1·10). The kernel of (1·10) can be expanded into power series of q:

$$\log\left|\frac{\vartheta_1(q,s)}{\vartheta_1^2(q^2,s)}\right|=-\log 2q^{3/4}-\log\chi-(3-4\chi^2)q^2-\frac{3+8\chi^2+16\chi^4}{2}q^2-\cdots\cdots$$

where $$\chi=\sin\pi s=\sin\frac{\sigma-\theta}{2}.$$

Then (1·10) can be written

$$\vartheta(\sigma)=\frac{K}{\pi}\int_0^{2\pi}e^{-3\tau}\sin\vartheta\cdot N(\sigma-\theta)\,d\theta. \quad (1\cdot12)$$

$$N(\sigma-\theta)=\sum_{n=1}^{\infty}P_n\cos n(\sigma-\theta)$$

$$P_1=1-2q^2+2q^4-\frac{4}{3}q^6+\cdots\cdots$$

$$P_2=\frac{1}{2}-q^4+O(q^8)$$

$$P_3=\frac{1}{3}-\frac{4}{3}q^6+\cdots\cdots$$

§ 2. Solution of the Integral Equations.

If we put $c_0=\sqrt{\frac{\lambda g}{2\pi}}$, c_0 means a wave velocity in case of infinitesimal amplitude and infinite depth of canal. $K=\frac{\lambda g}{2\pi c^2}=\frac{c_0^2}{c^2}$ is then more or less different from $1/P_1$, so we put $K=\frac{1}{P_1}-\varepsilon$. $\sqrt{P_1}\,c_0$ means wave velocity in case of finite amplitude and infinite depth.

Assuming $\varepsilon>0$, (1·12) becomes

$$\vartheta(\sigma)-\frac{1}{P_1\pi}\int_0^{2\pi}\vartheta(\theta)N(\sigma-\theta)\,d\theta$$
$$=-\frac{\varepsilon}{\pi}\int_0^{2\pi}\vartheta(\theta)N(\sigma-\theta)\,d\theta$$
$$+\frac{K}{\pi}\int_0^{2\pi}(e^{-3\tau}\sin\vartheta-\vartheta)N(\sigma-\theta)\,d\theta. \quad (2\cdot1)$$

We assume that τ and ϑ are comparatively small quantities and expand $e^{-3\tau}\sin\vartheta$ to the fourth order of ϑ and τ.

$$e^{-3\tau}\sin\vartheta=\vartheta-3\tau\vartheta+\frac{9}{2}\tau^2\vartheta-\frac{1}{6}\vartheta^3$$
$$-\frac{9}{2}\tau^3\vartheta+\frac{1}{2}\vartheta^3\tau+\cdots\cdots$$

So $e^{-3\tau}\sin\vartheta-\vartheta$ is small, right-hand side of (2·1) is small compared with the left-hand side. Putting zero the right-hand side of (2·1) as a first approximation, we obtain $\vartheta(\sigma)=\sin\sigma$, because

$$\int_0^{2\pi}\sin\theta N(\sigma-\theta)\,d\theta$$
$$=\int_0^{2\pi}\sum_{n=1}^{\infty}P_n\sin\theta\cos n(\sigma-\theta)\,d\theta$$
$$=P_1\int_0^{2\pi}\sin^2\theta\,d\theta=P_1\pi.$$

Next, putting $\vartheta=\sin\theta$ in the left-hand side of (2·1), (we denote this $f(\sigma)$) more precise solution of $\vartheta(\sigma)$ can be obtained from the Fredholm type of integral equation. In this case, according to the Schmidt's theorem,[4] $\int_0^{2\pi}f(\theta)\sin\theta\,d\theta=0$ must be realized in order to get solutions, $\sin\theta$ being considered an eigenfunction corresponding to an eigenvalue $1/P_1\pi$ of equation (2·1). However, $f(\theta)$ involves $\sin\theta$, then

$$\int_0^{2\pi}f(\theta)\sin\theta\,d\theta$$

can not be null. E. Schmidt has given so-called " die Methode de Kernzerspaltung "[4] in this trivial case, namely, a new kernel $M(\sigma-\theta)$ was given such that,

$$M(\sigma-\theta)=N(\sigma-\theta)-P_1\cos(\sigma-\theta)$$

Then, from (2.1) follows

$$\vartheta(\sigma)-\frac{1}{P_1\pi}\int_0^{2\pi}\vartheta(\theta)\mu(\sigma-\theta)\,d\theta=\mu\sin\sigma$$
$$+\frac{1}{P_1\pi}\int_0^{2\pi}(e^{-3\tau}\sin\vartheta-\vartheta)N(\sigma-\theta)\,d\theta$$
$$-\frac{\varepsilon}{\pi}\int_0^{2\pi}e^{-3\tau}\sin\vartheta N(\sigma-\theta)\,d\theta \quad (2\cdot2)$$

where $$\frac{1}{\pi}\int_0^{2\pi}\vartheta(\theta)\sin\theta\,d\theta=\mu \quad (2\cdot3)$$

and, $$\tau(\sigma)=K\int_0^{\sigma}e^{-3\tau}\sin\vartheta\,d\theta. \quad (2\cdot4)$$

μ is also considered a small quantity, and if a method of successive approximation is carried ϑ and τ could be expressed in a double power series of μ and ε.

Put $$\vartheta=a_1\mu+a_2\mu^2+a_3\mu\varepsilon+a_4\mu^3+a_5\mu^2\varepsilon$$
$$+a_6\mu\varepsilon^2+a_7\mu^4+a_8\mu^3\varepsilon+\cdots\cdots$$
$$\tau=b_1\mu+b_2\mu^2+b_3\mu\varepsilon+b_4\mu^3+b_5\mu^2\varepsilon$$
$$+b_6\mu\varepsilon^2+b_7\mu^4+b_8\mu^3\varepsilon+\cdots\cdots$$

then, coefficients, a_n, b_n are able to decided in order by putting (2·5) into (2·2) and (2·4), and equating the coefficients of $\mu^n\varepsilon^m$ (m and n are 0 or positive integers) on both member of formulae. For example, comparing the coefficient of μ^2 on both member of (2·2) we get,

$$a_2-\frac{1}{P_1\pi}\int_0^{2\pi}a_2\mu(\sigma-\theta)\,d\theta$$
$$=-\frac{1}{P_1\pi}\int_0^{2\pi}3a_1b_1N(\sigma-\theta)\,d\theta=\frac{3P_2}{2P_1^2}\sin2\sigma.$$

A method of success approximation is applied in this Fredholm's equation of second kind, and a_2 can be solved easily;

$$a_2=\frac{3P_2}{2P_1^2}\cdot\frac{P_1}{P_1-P_2}\sin 2\sigma.$$

From (2·4) we get,

$$b_2=\frac{1}{P_1}\int^{\sigma}(a_2-3a_1b_1)\,d\theta=-\frac{3}{4}\frac{\cos 2\sigma}{P_1(P_1-P_2)}+c_2.$$

c_2 being arbitrary constant involved into β_0 refered to in §1.

Now,

$$w=\frac{d}{dz}(\varphi+i\psi),\quad w=ce^{-i\omega},\quad \zeta=e^{2\pi i/\lambda}(\varphi+i\psi)$$

then

$$dz=\frac{\lambda}{2\pi i}e^{i\omega}\frac{d\zeta}{\zeta}.$$

On the free surface $\zeta=e^{i\sigma}$ and $z=\frac{\lambda}{2\pi}\int_0^{\sigma}e^{i\omega}d\theta$, hence,

$$x=\frac{\lambda}{2\pi}\int_0^{\sigma}e^{-\tau}\sin\vartheta\,d\theta,\quad y=\frac{\lambda}{2\pi}\int_0^{\sigma}e^{-\tau}\sin\vartheta\,d\theta. \tag{2·6}$$

Profiles of waves are decided from these formulae. Integrands involved in these expressions are expanded in series of ϑ and τ. Arbitrary constants involved in τ do not make $\int_0^{2\pi}e^{-\tau}\cos\vartheta\,d\theta$ just 2π, but in the first expression of (2·6), x must be λ when $\sigma=2\pi$. Therefore these arbitrary constants should be decided so that $\int_0^{2\pi}e^{-\tau}\cos\vartheta\,d\theta$ takes a value 2π. For example, c_2 should be $\frac{\lambda}{4P_1^2}(1-P_1^2)$ so that the coefficient of μ^2 involved in x is zero.

After cumbersome computation, a_n and b_n are calculated in terms of the sine and cosine, namely,

$$\left.\begin{aligned}
&a_1=\sin\sigma\\
&a_2=A_2\sin 2\sigma\\
&a_3=A_3\sin\sigma\\
&a_4=A_4\sin\sigma+A_4'\sin 3\sigma\\
&a_5=A_5\sin 2\sigma\\
&a_6=A_6\sin\sigma\\
&a_7=A_7\sin 2\sigma+A_7'\sin 4\sigma\\
&a_8=A_8\sin\sigma+A_8'\sin 3\sigma\\
&\quad\cdots\cdots\cdots\cdots\\
&b_1=B_1\cos\sigma\\
&b_2=B_2\cos 2\sigma+B_2'\\
&b_3=B_3\cos\sigma\\
&b_4=B_4\cos\sigma+B_4'\cos 3\sigma\\
&b_5=B_5\cos 2\sigma+B_5'\\
&b_6=B_6\cos\sigma\\
&b_7=B_7\cos 2\sigma+B_7'\cos 4\sigma'+B_7''\\
&b_8=B_8\cos\sigma+B_8'\cos 3\sigma\\
&\quad\cdots\cdots\cdots\cdots
\end{aligned}\right\}\tag{2·7}$$

where A_2, B_1, A_3, etc. are rational functions of $P_1, P_2, P_3, \cdots\cdots$ and can be expanded in power series of q.

$$\left.\begin{aligned}
&A_1=1\\
&A_2=\frac{3}{2}(1+6q^2+24q^4+\cdots\cdots)\\
&A_3=-(1-2q^2+2q^4-\cdots\cdots)\\
&A_4=1+6q^2+\frac{171}{4}q^4+\cdots\cdots\\
&A_4'=\frac{17}{6}+\frac{125}{6}q^2+\frac{2485}{24}q^4+\cdots\cdots\\
&A_5=-\left(\frac{15}{2}+36q^2+96q^4+\cdots\cdots\right)\\
&A_6=1-4q^2+8q^4+\cdots\cdots\\
&A_7=\frac{17}{2}+\frac{286}{3}q^2+\frac{2939}{3}q^4+\cdots\cdots\\
&A_7'=\frac{71}{12}+71q^2+\frac{20605}{36}q^4+\cdots\cdots\\
&A_8=-\left(8+\frac{97}{2}q^2+\frac{439}{2}q^4+\cdots\cdots\right)\\
&A_8'=-\left(\frac{87}{4}+\frac{4153}{24}q^2+\frac{10119}{4}q^4+\cdots\cdots\right)\\
&\quad\cdots\cdots\cdots\cdots\\
&B_1=-(1+2q^2+2q^4+\cdots\cdots)\\
&B_2=-\frac{3}{2}(1+6q^2+26q^4+\cdots\cdots)\\
&B_2'=q^2+2q^4+\cdots\cdots\\
&B_3=2\\
&B_4=-2\left(1+8q^2+\frac{227}{4}q^4+\cdots\cdots\right)\\
&B_4'=-\left(\frac{17}{6}+\frac{443}{18}q^2+\frac{9455}{72}q^4+\cdots\cdots\right)\\
&B_5=\frac{15}{2}+36q^2+111q^4+\cdots\cdots\\
&B_5'=-\left(\frac{1}{2}+3q^2+q^4+\cdots\cdots\right)\\
&B_6=-2(1-2q^2+2q^4+\cdots\cdots)\\
&B_7=-\left(\frac{17}{2}+\frac{572}{6}q^2+\frac{3041}{6}q^4+\cdots\cdots\right)\\
&B_7'=-\left(\frac{71}{12}+71q^2+\frac{20605}{36}q^4+\cdots\cdots\right)\\
&B_7''=\frac{1}{2}+4q^2+\frac{139}{4}q^4+\cdots\cdots\\
&B_8=-2\left(8+\frac{129}{2}q^2+\frac{665}{2}q^4+\cdots\cdots\right)\\
&B_8'=-\left(\frac{87}{4}+\frac{4153}{24}q^2+\frac{10119}{4}q^4+\cdots\cdots\right)
\end{aligned}\right\}\tag{2·8}$$

§3. Profiles and Velocity of Waves

In (2·6), $e^{-\tau}\cos\vartheta$ and $e^{-\tau}\sin\vartheta$ are expanded in series of τ and ϑ up to the fourth order:

$$\left.\begin{aligned} x&=\frac{\lambda}{2\pi}\int_0^{\sigma}\Big(1-\tau+\frac{\tau^2}{2}-\frac{\vartheta^2}{2}+\frac{\tau\vartheta^2}{2}-\frac{\tau^3}{6}\\ &\quad+\frac{\tau^4}{24}+\frac{\vartheta^4}{24}-\frac{\tau^2\vartheta^2}{4}+\cdots\cdots\Big)d\theta\\ y&=\frac{\lambda}{2\pi}\int_0^{\sigma}\Big(\vartheta-\tau\vartheta+\frac{\tau^2\vartheta}{2}-\frac{\vartheta^3}{6}-\frac{\tau^3\vartheta}{6}\\ &\quad+\frac{\tau\vartheta^3}{6}+\cdots\cdots\Big)d\theta. \end{aligned}\right\}\quad(3\cdot1)$$

Using (2·7) and (2·5), solutions of ϑ and τ are substituted into these equations and we obtain:

$$\left.\begin{aligned} \frac{2\pi x}{\lambda}&=\sigma+\Big(\frac{\mu}{P_1}-2\mu\varepsilon+S_4\mu^3+\cdots\cdots\Big)\sin\sigma\\ &\quad+\frac{1}{2}(S_2\mu^2+S_5\mu^2\varepsilon+S_7\mu^4+\cdots\cdots)\\ &\quad\times\sin 2\sigma+\frac{S_4'}{3}\mu^3\sin 3\sigma\\ &\quad+\frac{1}{4}(S_7'\mu^4+\cdots\cdots)\sin 4\sigma+\cdots\cdots\\ \frac{2\pi y}{\lambda}&=(\mu-P_1\mu\varepsilon+T_4\mu^3+\cdots\cdots)\\ &\quad\times(1-\cos\sigma)\\ &\quad+\frac{1}{2}(T_2\mu^2+T_5\mu^2\varepsilon+T_7\mu^4+\cdots\cdots)\\ &\quad\times(1-\cos 2\sigma)\\ &\quad+\frac{1}{3}(T_4'\mu^3+\cdots\cdots)(1-\cos 3\sigma)\\ &\quad+\frac{1}{4}(T_7'\mu^4+\cdots\cdots)(1-\cos 4\sigma)\\ &\quad+\cdots\cdots \end{aligned}\right\}\quad(3\cdot2)$$

where S_2, S_4, S_4', S_5, S_7, S_7'; T_2, T_4, T_4', T_5, T_7 and T_7' are expressed by integral functions of A_n and B_n, consequently expanded in power series with respect of q.

However, from (2·3) and (2·5) we obtain,

$$\mu+A_3\mu\varepsilon+A_4\mu^4+A_6\mu\varepsilon^2+A_8\mu^3\varepsilon+\cdots\cdots=\mu.$$

Solving ε with respect to μ,

$$\begin{aligned}\varepsilon&=-\frac{A_4}{A_3}\mu^2+\Big(\frac{A_4A_8}{A_3^2}-\frac{A_6A_4^2}{A_3^3}\Big)\mu^4+\cdots\cdots\\ &=\Big(1+8q^2+\frac{227}{4}q^4+\cdots\cdots\Big)\mu^2\\ &\quad-\Big(7+\frac{229}{2}q^2+1173q^4+\cdots\cdots\Big)\mu^4+\cdots\cdots\end{aligned}\quad(3\cdot3)$$

Substituting (3·3) into (3·2) we obtain finally,

$$\begin{aligned} \frac{2\pi x}{\lambda}&=\sigma+[(1+2q^2+2q^4+\cdots\cdots)\mu\\ &\quad+(q^2+12q^4+\cdots\cdots)\mu^3+\cdots\cdots]\sin\sigma\\ &\quad+\Big[\Big(1+5q^2+\frac{41}{2}q^4+\cdots\cdots\Big)\mu^2\\ &\quad+\frac{1}{2}\Big(1+\frac{1147}{36}q^2-\frac{1423}{4}q^4+\cdots\cdots\Big)\mu^4\\ &\quad+\cdots\cdots\Big]\sin 2\sigma+\Big[\Big(\frac{3}{2}+\frac{641}{54}q^2\\ &\quad+\frac{12983}{216}q^4+\cdots\cdots\Big)\mu^3+\cdots\cdots\Big]\sin 3\sigma\\ &\quad+\Big[\Big(\frac{8}{3}+\frac{952}{36}q^2+\frac{84887}{288}q^4+\cdots\cdots\Big)q^4\\ &\quad+\cdots\cdots\Big]\sin 4\sigma+\cdots\cdots\\ \frac{2\pi y}{\lambda}&=[\mu+(q^2+8q^4+\cdots\cdots)\mu^3+\cdots\cdots](1-\cos\sigma)\\ &\quad+\Big[\Big(1+5q^2+\frac{37}{2}q^4+\cdots\cdots\Big)\mu^2\\ &\quad+\frac{1}{2}\Big(1-\frac{1}{18}q^2+\frac{13139}{72}q^4+\cdots\cdots\Big)\mu^4\\ &\quad+\cdots\cdots\Big](1-\cos 2\sigma)+\Big[\Big(\frac{3}{2}+\frac{191}{18}q^2\\ &\quad+\frac{3661}{72}q^4+\cdots\cdots\Big)\mu^3+\cdots\cdots\Big](1-\cos 3\sigma)\\ &\quad+\Big[\Big(\frac{8}{3}+\frac{523}{18}q^2+\frac{20525}{96}q^4+\cdots\cdots\Big)\mu^4\\ &\quad+\cdots\cdots\Big](1-\cos 4\sigma)+\cdots\cdots \end{aligned}\quad(3\cdot4)$$

Wave height H', a distance between wave crests and mean water surface, is expressed by

$$H'=\frac{1}{\lambda}\int_{-\lambda/2}^{\lambda/2}y\,dx=\frac{1}{\lambda}\int_{-\pi}^{\pi}y(\sigma)\frac{dx}{d\sigma}d\sigma.\quad(3\cdot5)$$

Inserting (3·4) into this equation, we can express $\frac{2\pi H'}{\lambda}$ in series of μ. Then,

$$\begin{aligned}\frac{2\pi H'}{\lambda}&=\mu+\frac{1}{2}(1+8q^2+35q^4+\cdots\cdots)\mu^2\\ &\quad+\frac{1}{2}\Big(3+\frac{209}{9}q^2+\frac{4237}{36}q^4+\cdots\cdots\Big)\mu^3\end{aligned}$$

$$+\left(\frac{13}{6}+\frac{649}{36}\,q^2+\frac{66253}{288}\,q^4+\cdots\cdots\right)\mu^4$$
$$+\cdots\cdots \qquad (3\cdot6)$$

When a mean water depth is known, profiles of waves are drawn by means of (3·4). The wave velocity c is given from $\varepsilon=\frac{1}{P_1}-\frac{\lambda g}{2\pi c^2}$, (3·3) and (3·6);

$$\frac{c^2}{c_0^2}=(1-2q^2+2q^4+\cdots\cdots)$$
$$+\left(1+4q^2+\frac{131}{4}\,q^4+\cdots\cdots\right)\left(\frac{2\pi H'}{\lambda}\right)^2$$
$$-\left(1+12q^2+\frac{399}{4}\,q^4+\cdots\cdots\right)\left(\frac{2\pi H'}{\lambda}\right)^3$$
$$-\frac{31}{4}\left(1+\frac{3122}{279}\,q^2+\frac{95207}{1116}\,q^4+\cdots\cdots\right)$$
$$\times\left(\frac{2\pi H'}{\lambda}\right)^4+\cdots\cdots \qquad (3\cdot7)$$

Levi-Civita has obtained a formula giving wave velocity in case of infinite depth, namely

$$\frac{c^2}{c_0^2}=1+\left(\frac{2\pi H'}{\lambda}\right)^2-\left(\frac{2\pi H'}{\lambda}\right)^3$$
$$+\frac{7}{4}\left(\frac{2\pi H'}{\lambda}\right)^4-\cdots\cdots \qquad (3\cdot8)$$

(3·7) agrees with Levi-Civita's formula up to the third order of $\frac{2\pi H'}{\lambda}$ when $q=0$. Struik[5] has also studied the same problem as this paper in different manner, and attained a formula giving wave velocity such as,

$$\frac{c^2}{c_0^2}=\tanh\frac{2\pi H}{\lambda}$$
$$\times\left\{1-\frac{S_4+11S_2-6}{2D_1{}^6}\left(\frac{2\pi H'}{\lambda}\right)^2-\cdots\cdots\right\} \qquad (3\cdot9)$$

where $S_n=q^n+q^{-n}$, $D_n=q^{-n}-q^n$ or expanding in powers of q,

$$\frac{c^2}{c_0^2}=(1-2q^2+2q^4-\cdots\cdots)$$
$$-\frac{1}{2}\,(q^2+15q^4+\cdots\cdots)\left(\frac{2\pi H'}{\lambda}\right)^2-\cdots\cdots \qquad (3.10)$$

When $q=0$, this agrees with Levi-Civita's formula only by the first term of the right member.

In the case of small amplitude, the wave velocity is given by

$$c_m{}^2=\frac{\lambda g}{2\pi}\tanh\frac{2\pi H}{\lambda}=c_0{}^2\tanh\frac{2\pi H}{\lambda}$$

therefore from (3·9) $c<c_m$, but in reality when the wave height is finite, the wave velocity must be greater than that of small amplitude for the same value of λ and H. Therefore, Struik's formula is not in good agreement with this fact. But our formula is plausible because the wave velocity given by (3·7) is greater than c_m.

To compare with observations, a tank experiment of wave motion carried by Transportation Technical Research Institute served as a reference.[6] Tank dimension is 22 m long, 60 cm wide and 55 cm deep and the depth of mean water level is 30 cm. Waves generated in this water way has an irrotational and permanent character and whose amplitude is finite. Among these observations two cases were used here i.e.

	case 1	case 2
wave length	$\lambda=110$ cm	$\lambda=160$ cm
wave height	$A=6.46$ cm	$A=8.7$ cm
period	$T=0.87$ sec	$T=1.11$ sec
wave velocity	$c=126.4$ cm/sec	$c=144.1$ cm/sec

The second example gives $\frac{2\pi A}{\lambda}=0.37$, and $\frac{2\pi H}{\lambda}=1.71$, so $q=e^{-2\pi H/\lambda}=0.181$. When $\sigma=\pi$ is inserted in the second formula of (3·4) y becomes A, therefore

$$0.37=2\mu+0.084\mu^3+3.9\mu^3+\cdots\cdots$$

Solving this $\mu\fallingdotseq0.17$, and by means of (3.6) $\frac{2\pi H'}{\lambda}=0.21$. Wave velocity given by (3.7) is $\frac{c^2}{c_0^2}=0.957$. However, $c_0=\sqrt{\frac{\lambda g}{2\pi}}=131$ cm/sec, therefore $c\fallingdotseq128.4$ cm/sec.

The another example was treated in the same way. (Table 1)

Table 1.

		velocity of waves	
		case 1	case 2
observation		cm/sec 144.1	cm/sec 126.4
computed from	(3·7) in this paper	145.4	128.4
	$\sqrt{\frac{\lambda g}{2\pi}\tanh\frac{2\pi H}{\lambda}}$	143.7	126.8
	Struik's formula	143.2	126.7

Observed velocity must be greater than that of computed from $\sqrt{\frac{\lambda g}{2\pi}\tanh\frac{2\pi H}{\lambda}}$, because the amplitude of waves is finite, but the first observation contradicts with this. Theoretically, the velocity computed from Struik's formula must be greater than that of computed from $\sqrt{\frac{\lambda g}{2\pi}\tanh\frac{2\pi H}{\lambda}}$, because the finite wave height is treated in the former, but the infinitesimal wave height in the latter. Struik's formula fails in this respect. Our formula (3.7) does not fail, but observed velocity are smaller than expected. This is because of the complicated condition of the experiment, for example, reflected waves and the friction of water along the side of the tank will make wave velocity damped.

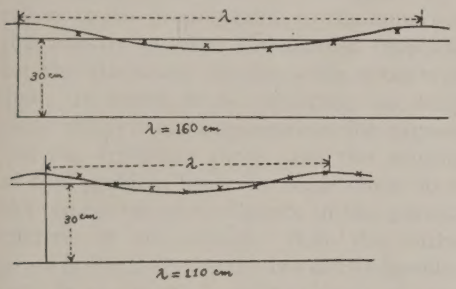

Fig. 4. Theoretical wave-profiles.
× points on the surface of observed waves.

Profiles of waves are drawn by means of (3·4) in each case. The theoretical profile and the observed profile resemble rather fairly. (Fig. 4)

References

(1) Milne-Thomson: Theoretical hydrodynamics, 1949, p. 303.
(2) T. Levi-Civita: Détermination rigoureuse des ondes permanents d'ampleur finie. Math. Annalen. Vol. 93, 1925.
(3) H. Villat: Leçons sur l'hydrodynamique. (1929) p. 56.
(4) E. Schmidt: Über die Auflösung der nichtlinearen Integral-gleichungen. Math. Annalen Vol. 65 (1908).
(5) K. J. Struik: Détermination rigoureuse des ondes irrotationelles periodiques dans un canal à profondeur finie. Math. Annalen Vol. 95. (1926).
(6) T. Hamada: Breakers and Beach erosions. Report of Transportation Technical Research Institute, Report No. 1 (1951).

A Short Note on the Dissipation of Seiche*

By

Takashi ICHIYE**

§1. The frictional force for the periodic current such as the secondary oscillation, tidal currents and seiche has been heretoforce put to be proportional to the velocity itself, owing chiefly to the simplicity of the mathematical treatment. However, it is taken to be proportional to the square of the velocity in other phenomena, especially for the stationary current, such as the river flow. It seems to be important to determine which is more appropriate for expressing the frictional force, and the solution of this problem leads to some clues about the production of turbulence in the periodic current of the nature. Here the author gives preliminary discussions of this problem only from the mathematical point of view.

§2. We take the simple case of one dimensional oscillation within the basin with closed ends, then the equation of motion with the friction is

$$\frac{\partial u}{\partial t}=-g\frac{\partial \zeta}{\partial x}-k|u|^n u \qquad (1)$$

where $-k|u|^n u$ is the friction term and $|u|$ means the absolute value of u. The equation of continuity is

$$\frac{\partial \zeta}{\partial t}=-h\frac{\partial u}{\partial x} \qquad (2)$$

where h is the depth. The friction term is here taken generally to be proportional to the $(n+1)$th power of the velocity. When we concern only of the principal oscillation, we can obtain some simple relation of the decrease of the amplitude with time under the assumption that the frictional force is small compared with the inertia term and restoring force. Only when the friction is proportional to the linear velocity we have the exact mathematical solution as easily seen from the above equations, since in this case we have

$$\zeta=A\cos\frac{\pi}{a}x,\quad u=-\frac{a}{\pi h}\dot{A}\sin\frac{\pi}{a}x \qquad (3)$$

from the eq. (2) and the boundary condition at the ends $x=0$ and $x=a$. Then we put (3) into (1) and we have exactly

$$\frac{d^2A}{dt^2}+gh\left(\frac{\pi}{a}\right)^2 A+k\dot{A}=0 \qquad (4)$$

and thus the solution is given easily. In general case, the normal mode of the oscillation cannot given by the simple form as (3) and the exact solution should be expressed by the sum of the series of $A_m\cos\frac{m\pi}{a}x$, which causes inevitably higher modes of oscillation. However, when the frictional force is slight, the normal function seems to be not so different from those given by (3). Further we assume that the relation (1) is valid only within the mean condition over the whole domaine and we take the normal functions given by (3), then we have

$$\frac{\partial}{\partial t}\int_0^a u\,dx=-g(\zeta_a-\zeta_0)-k\int_0^a u^n u\,dx$$

or

$$\frac{d^2}{dt^2}A+\omega^2 A+\frac{1}{2}k\left(\frac{a}{h\pi}\right)^n\frac{dA}{dt}^n\frac{dA}{dt}\times\int_0^\pi\sin^{n+1}\psi\,d\psi \qquad (5)$$

where $\omega^2=gh\left(\frac{\pi}{a}\right)^2$.

For the linear frictional force this gives the exact equation for the amplitude as seen easily, and for the square velocity case we have the frictional term as

$$\frac{1}{4}k\left(\frac{a}{h}\right)\dot{A}\,A=\epsilon\,\dot{A}\,A.$$

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The general solution of (5) is quite complicated but under the assumption that $\epsilon|\dot{A}|A$ is small compared with the term $\omega^2 A$, we can obtain the approximate solution of slowly varying amplitude and phase lag such as[1]

$$A = A(t)\sin[\omega t + \varphi(t)].$$

The amplitude for the square velocity case should be given by

$$\dot{A} = -\frac{\epsilon A^2 \omega}{2\pi}\int_0^{2\pi} |\cos\psi|\cos^2\psi\, dx$$

or

$$\dot{A} = A_0/(1+A_0 qt), \quad q = \frac{k}{2}\sqrt{g/h}. \qquad (6)$$

The general case is also given by

$$\dot{A} = -\frac{\epsilon_n A^{n+1}\omega^n}{2\pi}\int_0^{2\pi} |\cos^n\psi|\cos^2\psi\, d\psi,$$

$$\epsilon_n = \frac{k}{2}\left(\frac{a}{h\pi}\right)^n \int_0^{\pi}\sin^{n+1}\psi\, d\psi$$

or
$$A^{-n} - A_0^{-n} = nq_n t \qquad (7)$$

where we put q_n as $ckg^{1/2}h^{1/2-n}$ and c is the numerical constant easily determined. For the linear velocity case this method gives the exact decrement of the amplitude with time $A(t) = A_0 e^{-(k/2)t}$ as easily computed. The change of the phase lag with time can be also obtained from the simple calculation but for the practical purpose, the determination of the change of the phase lag or the period from the practical record is more uncertain than that of the amplitude, therefore the explicit expression of it is retained here.

The exact solution for the squre velocity case by use of the Fourier expansion will also be given in another paper. The above relations (6) and (7) indicate us that the decrement of the amplitude with time is dependent on the depth h and the initial amplitude A_0 for $n \neq 0$, whereas it is only dependent on A_0 for the frictional force proportional to the velocity itself. Such conditions suggests the possibility of the check from the practical observations and experiments about the question whether the frictional force is proportional to the velocity itself or not.

(1) Maclachlan: Ordinary non-linear differential equation. pp. 87~88. 1950. London.

Note on the Convective Circulation* (III)

By

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Abstract: On taking the variation of the Coriolis parameter into account and neglecting the non-linear terms, a convective current in an enclosed basin is discussed. Special attentions are paid to the velocity field near the eastern and western boundaries. A strong poleward component of the current is found near both the boundaries in the lower layer except in the surface layer where the existence of this strong component can not be concluded and a comparably strong eastward current takes place in the south equatorial region.

Introduction

The previous studies (Takano, 1955 a, b, c) except an example assuming far larger values for the coefficients of horizontal and vertical mixings than had been ordinarily taken, have been restricted to the central portion of a basin where there is no need for consideration of the longitudinal variation in the current velocity. The present part deals with the horizontal velocity in the vicinity of the boundaries. After the derivation of a solution including the wind stresses, we obtain a simpler approximate solution in order to easify to give a numerical example, assuming the vertical distribution of the current to be prescribed by the horizontal pressure gradient. This assumption may be valid for the eastward component of the velocity through the whole depth but not for the northward component at the surface layer, as inferred from the previous parts.

Equations and Solution

Putting

$$u=\sin\frac{\pi}{l}(y-y_0)\sum u_s\sin nx,$$

$$v=\sin\frac{\pi}{l}(y-y_0)\sum v_s\sin nx,$$

$$\rho=-\frac{4\Delta\rho}{L}(\rho_1\sinh\gamma z+\rho_2\cosh\gamma z)\cos\frac{\pi}{l}(y-y_0)\times\sum\frac{\sin nx}{n}+\frac{\rho_H-\rho_{-H}}{2H}z+\frac{\rho_H+\rho_{-H}}{2}$$

and the wind stress

$$\tau=\sin\frac{\pi}{l}(y-y_0)\sum\tau_s\sin nx$$

with the same co-ordinate system and the notations as in the part I, neglecting the non-linear terms and daring to replace $A_h\frac{\partial^2 u}{\partial y^2}$ and $A_h\frac{\partial^2 v}{\partial y^2}$ by $-ku$ and $-kv$, respectively, though this replacement may be rather unreasonable, it follows that

$$\begin{aligned}u_s=&U_s(\rho_1\cosh\gamma z+\rho_2\sinh\gamma z)\\&+\frac{A_v}{f(\alpha^2+\beta^2)}\Big[(G\alpha-F\beta)\Big\{-\frac{\tau_s}{2A_v}\Big(\frac{A_hn^2+k}{A_v}-\alpha^2+\beta^2\Big)+U_s\gamma\rho_2\cosh\gamma H\cdot(\alpha^2-\beta^2-\gamma^2)\Big\}\\&+\alpha\beta(F\alpha+G\beta)\Big(-\frac{\tau_s}{A_v}-2U_s\gamma\rho_2\cosh\gamma H\Big)\\&-(K\alpha-H\beta)\Big\{-\frac{\tau_s}{2A_v}\Big(\frac{A_hn^2+k}{A_v}-\alpha^2+\beta^2\Big)\\&+U_s\gamma\rho_1(\gamma^2-\alpha^2+\beta^2)\sinh\gamma H\Big\}\\&-(H\alpha+K\beta)\alpha\beta\Big(-\frac{\tau_s}{A_v}+2U_s\gamma\rho_1\sinh\gamma H\Big)\Big]\\&+\frac{1}{H}\Big\{\Big(\frac{p_1^2+p_2^2-\gamma^2}{p_1^2p_2^2}-\frac{1}{\gamma^2}\Big)\gamma U_s\rho_1\sinh\gamma H\\&-\frac{\tau_s}{2A_vp_1^2p_2^2}\Big(p_1^2+p_2^2-\frac{A_hn^2+k}{A_v}\Big)\Big\},\end{aligned}\qquad(1)$$

$$\begin{aligned}fv_s=&U_s(\rho_1\cosh\gamma z+\rho_2\sinh\gamma z)(A_hn^2+k-A_v\gamma^2)\\&+\frac{A_v\{A_hn^2+k-A_v(\alpha^2-\beta^2)\}}{f(\alpha^2+\beta^2)}\Big[(G\alpha-F\beta)\\&\times\Big\{-\frac{\tau_s}{2A_v}\Big(\frac{A_hn^2+k}{A_v}-\alpha^2+\beta^2\Big)\\&+(\alpha^2-\beta^2-\gamma^2)U_s\gamma\rho_2\cosh\gamma H\Big\}\end{aligned}$$

* Received Aug. 3, 1955.

** Geophysical Institute, Tokyo University.

$$
\begin{aligned}
&+(F\alpha+G\beta)\,\alpha\beta\Big(-\frac{\tau_s}{A_v}-2U_s\gamma\rho_2\cosh\gamma H\Big)\\
&-(K\alpha-H\beta)\,\Big\{-\frac{\tau_s}{2A_v}\Big(\frac{A_hn^2+k}{A_v}-\alpha^2+\beta^2\Big)\\
&+(\gamma^2-\alpha^2+\beta^2)\,U_s\gamma\rho_1\sinh\gamma H\Big\}\\
&-(H\alpha+K\beta)\,\alpha\beta\Big(-\frac{\tau_s}{A_v}+2U_s\gamma\rho_1\sinh\gamma H\Big)\Big]\\
&-\frac{2A_v^2\alpha\beta}{f(\alpha^2+\beta^2)}\Big[(F\alpha+G\beta)\,\Big\{-\frac{\tau_s}{2A_v}\Big(\frac{A_hn^2+k}{A_v}\\
&-\alpha^2+\beta^2\Big)+(\alpha^2-\beta^2-\gamma^2)\,U_s\gamma\rho_2\cosh\gamma H\Big\}\\
&-(G\alpha-F\beta)\,\alpha\beta\Big(-\frac{\tau_s}{A_v}-U_s\gamma\rho_2\cosh\gamma H\Big)\\
&-(H\alpha+F\beta)\,\Big\{-\frac{\tau_s}{2A_v}\Big(\frac{A_hn^2+k}{A_v}-\alpha^2+\beta^2\Big)\\
&+(\gamma^2-\alpha^2+\beta^2)\,U_s\gamma\rho_1\sinh\gamma H\Big\}\\
&+(K\alpha-H\beta)\,\alpha\beta\Big(-\frac{\tau_s}{A_v}+U_s\gamma\rho_1\sinh\gamma H\Big)\Big]\\
&+\frac{A_hn^2+k}{H}\Big\{\Big(\frac{p_1^2+p_2^2-\gamma^2}{p_1^2p_2^2}-\frac{1}{\gamma^2}\Big)\\
&\times\gamma U_s\rho_1\sinh\gamma H\\
&-\frac{\tau_s}{2A_vp_1^2p_2^2}\Big(p_1^2+p_2^2-\frac{A_hn^2+k}{A_v}\Big)\Big\}+\frac{\tau_s}{2H},
\end{aligned}
\tag{2}
$$

with the following boundary conditions:

$$
\left.
\begin{aligned}
&u=v=0, && \text{at}\quad x=0,\ x=L\quad\text{(eastern and western boundaries)}\\
&A_v\frac{\partial u}{\partial z}=-\tau,\\
&A_v\frac{\partial v}{\partial z}=w=0, && \text{at}\quad z=-H\quad\text{(surface)}\\
&\rho=-\Delta\rho\cos\frac{\pi}{l}(y-y_0)+\rho_{-H},\\
&A_v\frac{\partial u}{\partial z}=A_v\frac{\partial v}{\partial z}=w=0,\\
&\rho=\rho_H, && \text{at}\quad z=H\quad\text{(bottom)}
\end{aligned}
\right\}
\tag{3}
$$

Here,

$$
\left.
\begin{aligned}
F&=\frac{\cos\beta(z+H)\sinh\alpha(z-H)+\cos\beta(z-H)\sinh\alpha(z+H)}{\cos 2\beta H+\cosh 2\alpha H},\\
G&=\frac{\sin\beta(z+H)\cosh\alpha(z-H)+\sin\beta(z-H)\cosh\alpha(z+H)}{\cos 2\beta H+\cosh 2\alpha H},\\
H&=\frac{\cos\beta(z-H)\sinh\alpha(z+H)+\cos\beta(z+H)\sinh\alpha(H-z)}{\cos 2\beta H+\cosh 2\alpha H},\\
K&=\frac{-\sin\beta(z+H)\cosh\alpha(z-H)+\sin\beta(z-H)\cosh\alpha(H+z)}{\cos 2\beta H+\cosh 2\alpha H}.
\end{aligned}
\right\}
\tag{4}
$$

The series thus obtained are only slowly convergent, then in order to avoid tedious calculations, the vertical distribution of the current is assumed to be provided by the pressure gradient only. This assumption weekens remarkably the northward component of the current at the surface layer compared with that obtained without this assumption.

Combined with the boundary conditions

$$
\left.
\begin{aligned}
&u=v=0, && \text{at}\quad x=0,\ x=L\quad\text{(eastern and western boundaries)}\\
&\rho=-\Delta\rho\cos\frac{\pi}{l}(y-y_0)+\rho_0, && \text{at}\quad z=0\quad\text{(surface)}\\
&\rho=\rho_H, && \text{at}\quad z=H\quad\text{(bottom)}\\
&w=0, && \text{at}\quad z=0,\ z=H,
\end{aligned}
\right\}
\tag{5}
$$

the equations of motion give

$$u=-\sin\frac{\pi}{l}(y-y_0)\Big(\coth\gamma H\cdot\cosh\gamma z-\sinh\gamma z-\frac{1}{\gamma H}\Big)U_p\Big[-1+\frac{1}{p_1^2-p_2^2}\times\Big\{-p_2^2(1-\cosh p_1L)\frac{\sinh p_1x}{\sinh p_1L}-p_2^2\cosh p_1x+p_1^2(1-\cosh p_2L)\frac{\sinh p_2x}{\sinh p_2L}+p_1^2\cosh p_2x\Big\}\Big], \tag{6}$$

$$v=\frac{U_p}{f}\sin\frac{\pi}{l}(y-y_0)\Big(\coth\gamma H\cdot\cosh\gamma z-\sinh\gamma z-\frac{1}{\gamma H}\Big)\Big[-(A_v\gamma^2-k)+\frac{A_v\gamma^2-k}{p_1^2-p_2^2}\times\Big\{-p_2^2(1-\cosh p_1L)\frac{\sinh p_1x}{\sinh p_1L}-p_2^2\cosh p_1x+p_1^2(1-\cosh p_2L)\frac{\sinh p_2x}{\sinh p_2L}+p_1^2\cosh p_2x\Big\}-A_hp_1^2\Big\{-p_2^2(1-\cosh p_1L)\times\frac{\sinh p_1x}{\sinh p_1L}-p_2^2\cosh p_2x\Big\}-A_hp_2^2\Big\{p_1^2(1-\cosh p_2L)\frac{\sinh p_2x}{\sinh p_2L}+p_1^2\cosh p_2x\Big\}\Big], \tag{7}$$

with the approximate equation of continuity

$$\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}+\frac{\partial w}{\partial z}=0, \tag{8}$$

where

$$p_1=\sqrt{\frac{k-A_v\gamma^2}{A_h}+\frac{if}{A_h}},\quad p_2=\sqrt{\frac{k-A_v\gamma^2}{A_h}-\frac{if}{A_h}}$$

and

$$U_p=-\frac{gfm\Delta\rho}{\gamma\{(A_v\gamma^2-k)^2+f^2\}}. \tag{9}$$

The above solutions (6) and (7) are transformed into

$$u=-U_p\sin\frac{\pi}{l}(y-y_0)\Big(\coth\gamma H\cosh\gamma z-\sinh\gamma z-\frac{1}{\gamma H}\Big)\Big(1-L+\frac{k-A_v\gamma^2}{f}M\Big), \tag{10}$$

$$v=\frac{U_p}{f}\sin\frac{\pi}{l}(y-y_0)\Big(\coth\gamma H\cosh\gamma z-\sinh\gamma z-\frac{1}{\gamma H}\Big)\times\{(A_v\gamma^2-k)(1-L)+fM\}. \tag{11}$$

Here L and M are

$$L=\frac{\cosh\alpha(x-L)\cos\beta x+\cosh\alpha x\cos\beta(x-L)}{\cos\beta L+\cosh\alpha L}, \tag{12}$$

$$M=\frac{\sin\beta x\sinh\alpha(x-L)+\sin\beta(x-L)\sinh\alpha x}{\cos\beta L+\cosh\alpha L}, \tag{13}$$

on putting $p_1=\alpha+\beta i$ and $p_2=\alpha-\beta i$.

If $k-A_v\gamma^2$ is negligibly small compared with $|f|$ and the small quantities are neglected, it follows that

$$u=\frac{g\pi\Delta\rho}{\gamma fl}\sin\frac{\pi}{l}(y-y_0)\times\Big(\coth\gamma H\cdot\cosh\gamma z-\sinh\gamma z-\frac{1}{\gamma H}\Big)\times\Big(1-e^{-\alpha x}\cos\beta x+\frac{A_v\gamma^2-k}{f}e^{-\alpha x}\sin\beta x\Big), \tag{14}$$

$$v=\frac{g\pi\Delta\rho}{\gamma fl}\sin\frac{\pi}{l}(y-y_0)\times\Big(\coth\gamma H\cdot\cosh\gamma z-\sinh\gamma z-\frac{1}{\gamma H}\Big)\times\Big\{-\frac{A_v\gamma^2-k}{f}(1-e^{-\alpha x}\cos\beta x)+e^{-\alpha x}\sin\beta x\Big\}. \tag{15}$$

These remain valid near the western boundary, provided x is regarded as the distance from the boundary.

Except in the close vicinity of the equator where the Coriolis parameter tends to be infinitesimal, $\frac{\partial v}{\partial x}$ vanishes with

$$x_0=\frac{\pi}{4|\beta|}=\frac{\pi}{2\sqrt{2}}\sqrt{\frac{A_h}{|f|}}. \tag{16}$$

Since the absolute value of the Coriolis parameter increases with the latitude, the maximum northward or southward component of the current appears in closer proximity to the boundary in higher latitudes.

When $A_v\gamma^2-k$ is sufficiently small in comparison with $|f|$, (14), (15) and (8) give, near the western boundary,

$$w=-\frac{g\pi\Delta\rho}{\gamma^2lf}e^{-\alpha x}\times\Big(\coth\gamma H\cdot\sinh\gamma z-\cosh\gamma z-\frac{z}{H}+1\Big)\times\Big[\frac{\pi}{l}\cos\frac{\pi}{l}(y-y_0)-\sin\frac{\pi}{l}(y-y_0)$$

$$\times\left\{\frac{df}{dy}\left(\frac{1}{f}\sin\alpha x+\frac{x}{2\sqrt{A_h f}}\sin\overline{\alpha x-\frac{\pi}{4}}\right)\right.$$
$$\left.\mp\sqrt{2}\,\alpha\sin\left(\alpha x+\frac{\pi}{4}\right)\right\}\Big] \qquad (17)$$

in the northern hemisphere and

$$w=-\frac{g\pi\Delta\rho}{\gamma^2 lf}\,e^{-\alpha x}$$
$$\times\left(\coth\gamma H\cdot\sinh\gamma z-\cosh\gamma z-\frac{z}{H}+1\right)$$
$$\times\Big[\frac{\pi}{l}\cos\frac{\pi}{l}(y-y_0)-\sin\frac{\pi}{l}(y-y_0)$$
$$\times\left\{-\frac{df}{dy}\left(\frac{1}{f}\sin\alpha x-\frac{x}{2\sqrt{A_h|f|}}\sin\overline{\alpha x-\frac{\pi}{4}}\right)\right.$$
$$\left.\mp\sqrt{2}\,\alpha\sin\left(\alpha x+\frac{\pi}{4}\right)\right\}\Big] \qquad (18)$$

in the southern hemisphere.

Here the positive sign refers to the neighborhood of the eastern boundary and in this case x is taken as the distance from the western boundary.

The figures 1 to 5 illustrate a numerical example of the solutions (14) and (15) in unit $10^4\Delta\rho\left(\coth\gamma H\cdot\cosh\gamma z-\sinh\gamma z-\frac{1}{\gamma H}\right)$ for

$A_h=0.5\times10^8$ g cm^{-1} sec^{-1}, $l=2\pi\times10^8$ cm,
$A_v\gamma^2-k\simeq0$, $A_v=0.5\times10^2$ g cm^{-1} sec^{-1},
$g=980$ cm sec^{-2}, $\omega=7.29\times10^{-5}$ sec^{-1}

and y_0 corresponding to the latitude 8°N, according to Wüst et al. (Wüst et al., 1954).

The variation of

$$Z=\coth\gamma H\cdot\cosh\gamma z-\sinh\gamma z-\frac{1}{\gamma H}$$

giving the vertical distribution of the current is tabulated in the Table 1 for the same numerical values.

Table 1.

Values of $Z=\coth\gamma H\cdot\cosh\gamma z-\sinh\gamma z-\frac{1}{\gamma H}$

z (m)	0	100	500	1000	2000	3000	4000
Z	0.537	0.488	0.317	0.148	-0.075	-0.190	-0.226

Conclusions

The northward component in higher latitudes than 8°N or the southward component in other latitudes has the maximum values at a distance from the boundaries decreasing with latitude and decreases progressively to vanishingly small values off the boundaries, owing to assumptions that there is no longitudinal variation in density and the vertical structure is prescribed by the pressure gradient only. If $\Delta\rho$ is taken to be 2.5×10^{-3} which is found in the Pacific Ocean, for instance, the maximum value in the 500 m layer at 20°N is about 3 cm sec^{-1}, though it is possible that the first assumption mentioned above does not remain valid there, because of the strong tonguelike intrusion of less dense water into this region in the upper layer and more dense water in the lower layer.

Without the second assumption, the northward component may not be vanishingly small at the surface layer in the central portion of the basin, having a value of the order of 10 to $100\times\Delta\rho$ as inferred from the part I, and it can not be necessarily concluded that this component is intensified correspondingly at the surface layer near the boundaries, but at least in the lower layer the intensification is reliable.

In the first approximation, the integration of the velocity component from the surface down to the bottom vanishes everywhere, according to the equation of continuity (8); however, rigorously speaking, the equation of continuity

$$\frac{\partial}{\partial x}(\rho u)+\frac{\partial}{\partial y}(\rho v)+\frac{\partial}{\partial z}(\rho w)$$
$$=A_h\left(\frac{\partial^2\rho}{\partial x^2}+\frac{\partial^2\rho}{\partial y^2}\right)+A_v\frac{\partial^2\rho}{\partial z^2}$$

leads to

$$\frac{\partial M_x}{\partial x}+\frac{\partial M_y}{\partial y}=A_h\left(\int_s^b\frac{\partial^2\rho}{\partial x^2}dz+\int_s^b\frac{\partial^2\rho}{\partial y^2}dz\right)$$
$$+A_v\frac{\partial\rho}{\partial z}\Big|_b-A_v\frac{\partial\rho}{\partial z}\Big|_s$$

after the integration from the surface to the bottom, where

$$M_x=\int_s^b\rho u\,dz \quad\text{and}\quad M_y=\int_s^b\rho v\,dz,$$

and the suffixes s and b refer to the surface and the bottom, respectively. Since the right-hand side does not vanish generally the mass transport can exist, even though

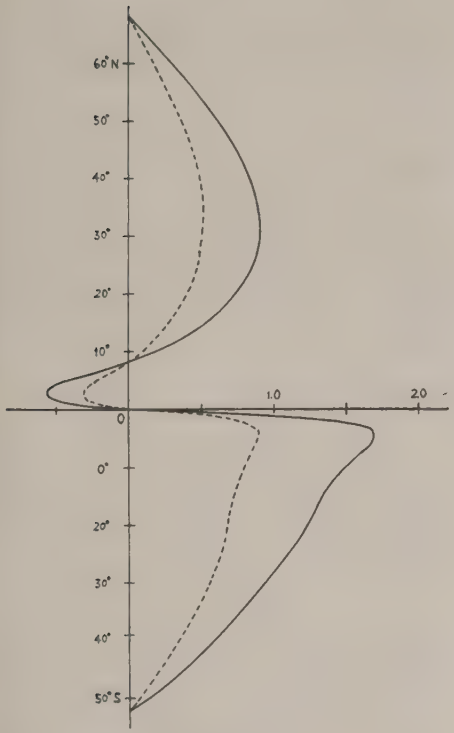

Fig. 1. Eastward component.
unit in
$10^4 \Delta\varphi(\coth h\gamma H \cdot \cos h\gamma z - \sin h\gamma z - 1/H\gamma)$.
$(\mathrm{cm\,sec^{-1}\,g^{-1}})$.
full curve for x=100km
broken curve for x=5km

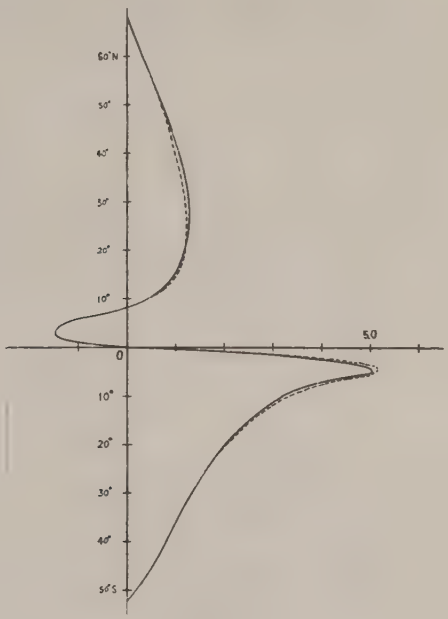

Fig. 2. Eastward component.
same unit.
full curve for x=100km
broken curve for x=50km

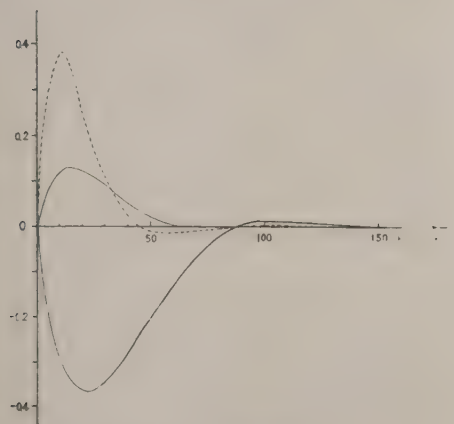

Fig. 3. Northward component.
same unit.
heavy curve at 5°N
thin curve at 10°N
broken curve at 20°N

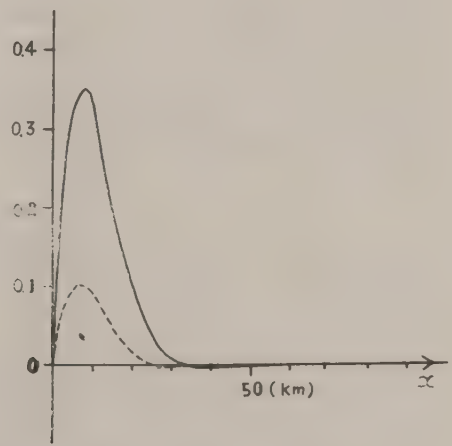

Fig. 4. Northward component.
same unit.
full curve at 40°N
broken curve at 60°N

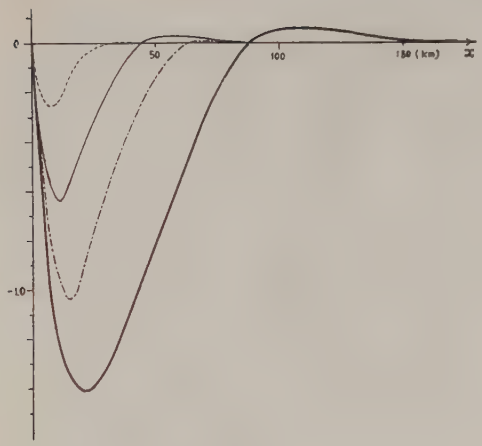

Fig. 5. Northward component.
same unit.
heavy curve at 5°S
curve at 10°S
thin curve 20°S
broken curve at 40°S

the stress at the bottom and the wind stress are not taken into account.

Such a pattern of the convective circulation is not always in agreement with the actual pattern of the ocean circulation. These disagreements may be attributed to the neglect of the wind stress, to the density distribution where the thermocline is not found anywhere, because of the constant vertical eddy diffusivity, to the exclusion of the non-linear terms in the equation of density diffusion and to the assumptions that the vertical structure of the current is determined by the pressure gradient only and there is no longitudinal variation in the density.

Of course, the boundary condition with respect to the density has not to be given by that at the surface, because the actual distribution of the surface density is the result of some effects including the wind actions and so the quantities such as the radiations, the evaporation, the precipitation etc. which cause the change of density have to be taken as the boundary condition at the surface. It seems to us, however, that there are no considerable differences between the two solutions derived from these different boundary conditions.

Acknowledgments

We wish to express our sincerest thanks to Dr. K. Hidaka who eucourged us during this study and to Dr. K. Yoshida as well as Mr. M. Tsuchiya and Mr. K. Ogawa for their valuable discussions.

References

Takano, Kenzo 1955 a: An example of Thermal Convective Current. Records of Oceanographic Works in Japan, Vol. 2, 76

———— 1955 b: Note on the convection circulation (I), ditto Vol. 2.

———— 1955 c: Note on the convection circulation (II), Journ. Oceanogr. Soc. Japan, Vol. 11, 39.

Wüst, G., Willi Brogmus und Eberhard Noodt; 1954: Die zonale Verteilung von Salzgehalt, Niederschlag, Verdunstung. Temperatur und Dichte an der Oberfläche der Ozeane. Kieler Meeresforschungen, Band X, 137.

An Example of Variations in Oceanic Circulation in Response to the Variations in Wind Field*

By

Kozo YOSHIDA**

Abstract: An example of seasonal variations in surface oceanic circulation is revealed from the data off the California coasts. An account for them may probably be given by a simple model in which the rearrangement of density through wind-induced vertical motion is essential to the changes. This mechanism might also apply more widely to annual variations in oceanic circulation in response to those in atmospheric pressure field. An emphasis may be placed on the point that the variations in currents are in quasi-stationary balance with pressure field in the ocean surface layer which varies as integral effect of winds. A possibility of quantitative prediction might be suggested.

Introduction

It should be a matter of great difficulty at present to attempt to deal with variations in oceanic circulation. This involves time constants of the ocean circulation which so far have not been well known. A lack of this knowledge is primarily ascribed to the difficulties in doing observations with sufficient density and simultaneity at sea. It is therefore required at first that more observations should be made in this direction as they are being activated in recent years. Nevertheless, it might be desirable to get some rough ideas of such variations, even ahead of observations, considering that any individual data which are collected to be analyzed must represent various particular phases of ever-changing pattern of the phenomena. In this connection, the author is inclined to feel that some of the foregoing theoretical approaches might be misleading(1) and might require discussions.

For an annual variation, for instance, the dynamical balance would approximately be quasi-stationary without acceleration, except for the boundary zone such as close to coasts. Thus the instantaneous field of surface water cirulation which varies season to season would be well represented as a geostrophic flow, nearly in phase with the corresponding instantaneous surface slope. However, this does not mean that the currents respond *to winds* quasi-steadily, since readjustment of pressure field to winds will require some considerable time length. Mass distributions are changing during a large part of the period when vertical motion is induced by winds, and thus the surface current varies as an integral effect of winds. The time lag may therefore be mainly due to the time required for mass readjustment and not due to the acceleration.

In the waters off the California coasts, the observations taken by serial cruises at the Scripps Institution of Oceanography appear to have confirmed the facts that northerly California Current develops inshore during periods of upwelling. In the absense of prevailing winds that cause upwelling, a countercurrent (northward) would appear on the coastal side. These features of seasonal changes in surface currents are *consistently* revealed by the preceding observations. Some examples

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(1) It seems somewhat misleading to apply the following simplified models to the variations of periods longer than several days:
  (1) The acceleration terms are formally included in the dynamical equations, whereas the changes in mass distributions are essentially disregarded or not critically taken into consideration;
  (2) Quasi-static equilbrium is assumed to apply to the variations in circulation patterns as well as in mass distributions.

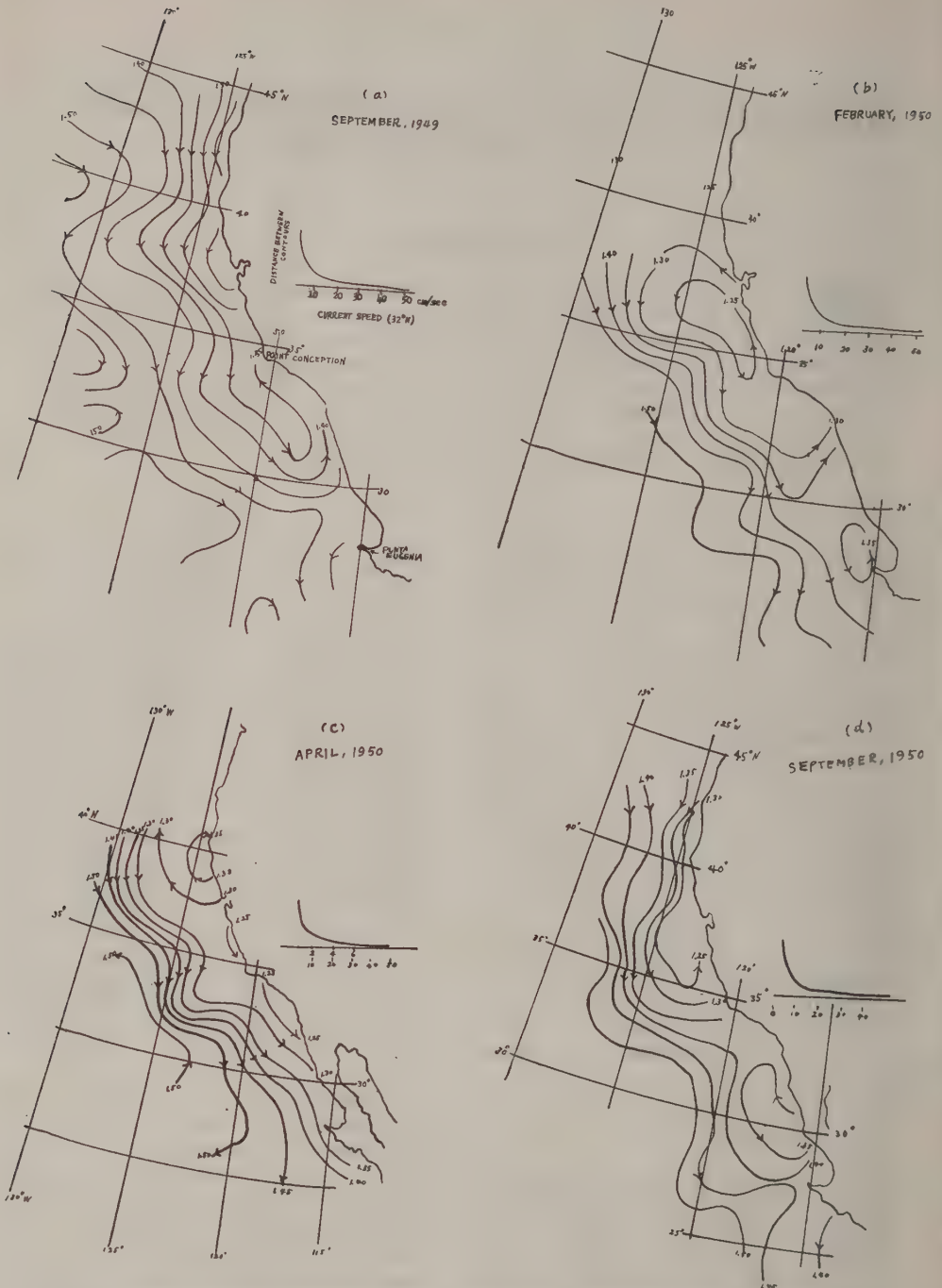

Fig. 1. Dynamic Height Anomalies (0 over 1000 decibars) : Contour interval is 0.05 dynamic meters, and the current direction is as indicated by arrows: (a) CCOFI Cruise 4909 (MLR, cruise 7) September 5 to September 18, 1949. (b) CCOFI Cruise 5002 (MLR, cruise 11) January 31 to February 19, 1950. (c) CCOFI Cruise 5004 (MLR, cruise 13) April 3 to 17, 1950. (d) CCOFI Cruise 5009 (MLR, cruise 18) September 6 to 22, 1950.

are illustrated in Figs. 1a to 1d, taken from recent CCOFI (California Cooperative Oceanic Fisheries Investigation) cruise results. The cruises in 1937 and in 1938 also reveal remarkably consistent pictures. Refer to those figures* prepared by Sverdrup and Allen (1939) and those by Sverdrup and Fleming (1941). It has also been indicated that seasonal occurrence of upwelling seems to be in phase with seasonal changes in areas of positive wind stress curl off the coast. (Yoshida et al, 1955). The annual variations in mean sea level along the coasts have been found to be closely related to those in the high pressure field over the North Pacific (Jacobs, 1939). Recently, Munk, Pattullo, Revelle and Strong have confirmed that the annual variations in mean sea level are in satisfactory agreement with those in density field under an isostatic adjustment within an upper 400 meter layer.

These observed facts which have been accumulated in the California waters appear to support the idea outlined above at least in the cases of this current system, and might further enable us to go into somewhat quantitative discussions on variations in the currents. A possibility of *forecasting* oceanographic conditions might be suggested. Not only to the California current system, but also to the larger-scale circulation system, the similar mechanism might apply.

A Model for California Current System

For simplicity, let us take x axis eastward, y axis northward and z axis vertically downward, and assume the coastline extends north to south. Except for the narrow zone close to the coast, some 50 km from coast, the dynamical equation for the upper mixed layer may be given by

$$-fv=-\partial p/\partial x \qquad (1)$$

$$fu=-\partial p/\partial y+\tau_y/H \qquad (2)$$

where u and v are eastward and northward component of velocity respectively, f is Coriolis parameter, p is pressure, τ_y is northward component of wind stress, H is a mean depth of upper mixed layer.

From the above equations, vertical velocity w_h at a thermocline ($z=h$) is seen, to a first approximation, to be

$$w_h\sim-(1/f)\,\partial\tau_y/\partial x \qquad (3)$$

provided that planetary vorticity $\beta\ v$, where $\beta=df/dy$, is neglected, as could be justified for the surface current in the areas concerned (Yoshida et al, 1955). From the condition of nearly isostatic balance:**

$$\partial p/\partial x\sim g\Delta\rho\,(\partial h/\partial x) \qquad (4)$$

where $\Delta\rho$ is a difference between the mean density in surface layer and in lower layer, and from the condition of mass conservation:

$$w_h\sim\partial h/\partial t \qquad (5)$$

it follows that the change in velocity component v is given by

$$\partial v/\partial t\sim(g\Delta\rho/f)\,(\partial w_h/\partial x)\sim-(g\Delta\rho/f^2)\,(\partial^2\tau_y/\partial x^2) \qquad (6)\ .$$

This eq. (6) implies that southward current will be accelerated during the period when upwelling takes place inshore. Combined this together with those features of seasonal variations in the wind field and in vertical motion off the California coasts which have been indicated by the author (Yoshida and Mao, 1955), it may be considered that the present model may represent observed features of the variations satisfactorily.

* Refer to
(1) Fig. 4 (Cruise 1, March 25–April 3, 1937), Fig. 6 (Cruise 3, June 22–July 2, 1937), in the paper by Sverdrup and Fleming (1941), and
(2) Fig. 39 (Cruise 2, April, 1938), Fig. 41 (Cruise 4, August, 1938), Fig. 43 (Cruise 6, December, 1938), in the paper by Sverdrup and Allen (1939).

** From the examinations of cruise data off the California coast, February to May, 1949, and of the wind conditions over the areas off the coasts, Defant estimated that the sea surface would adjust itself to the wind within *one or two days* and that this quasistationary state would be maintained as long as the wind continues. (Albert Defant, Memorandum concerning the work of the Marine Life Research Program, La Jolla, 1950).

As mentioned before, the seasonal changes in w occur quasisteadily as winds vary season to season, and will result in the seasonal changes in thickness of mixed layer which associate the changes in surface currents.(2) Now, quasi-static current field involves wind-induced vertical motion, which responds to winds without much delay, perhaps less than several days. The changes in surface currents are thus "caused" by non-uniformity in wind stress curl through the mechanism of vertical motion. The time lag for the response of the ocean surface circulation is due to the time required to establish a new mass field which itself is brought about by horizontal divergence in the basic currents. When winds vary season to season, for exmple, the surface currents vary with time lag of order of three months in absence of lateral diffusion. It appears, on the other hand, that the outstanding features in the variations revealed by observations off the California coasts may suggest the lag to be less, at most one to two months. The role of lateral diffusion may therefore be suggested, and (5) may be replaced, as previously derived (Yoshida, 1955) by

$$w_h \sim \partial h/\partial t - B\partial^2 h/\partial x^2, \qquad (7)$$

where B is a coefficient of lateral mixing. In fact, the second term indicating mixing will be quite comparable to the first when the annual variations are concerned. If we take $\sigma \sim 2\times10^{-7}$ (period : 1 year), $k \sim 6\times10^{-8}$ (wavelength : 10^3 km) and $B \sim 2\times10^7$, the phase lag will be

$$\tan^{-1}\frac{\sigma}{Bk^2} \sim 70^\circ \sim 2.3 \text{ months}$$

where σ and k denote frequency and wave number of the disturbance, respectively. Although the matter may critically depend upon a magnitude of B which is not well known, it appears that the time lag for the wave of significant annual range will perhaps be one to two months. The corresponding range of annual variations in v is found approximately to be 20 cm/sec, which may be in fair agreement with observed conditions encountered in the California waters. Typical examples of the features are clearly illustrated in Figs. 1a to 1d.

Within a narrow strip close to the coast, perhaps within 50 to 100 km from the coast, time changes in relative vorticity will be considerable (Yoshida, 1955), and lateral mixing will be essential to the variations probably due to large internal waves, horizontal eddies and large variability itself in the current, so that minor fluctuation in winds may result in relatively quick response of coastal currents which without much delay establish quasi-static equilibrium.

In lower layers of California waters, it appears to have been indicated that subsurface north-south transport changes from season to season nearly in phase with the corresponding changes in wind stress curl (Yoshida et al, 1955). The processes of adjustment might be represented by

$$\partial^2 v_2/\partial x \partial t + \beta v_2 = (f/H_2)\,\mathrm{curl}\,\tau$$

where v_2 is a meridional component of velocity in lower layer, and H_2 is a thickness of the lower layer.

Probably, in the current system where north-south component of velocity is predominant, the response of the current to winds is comparatively rapid, say, with a phase lag of order of a month or less. The response of the flow at depths greater than 500m will, however, be much slower.

Other Current Systems

The Kuroshio, Gulf Stream, and perhaps whole oceanic circulation may vary in somewhat similar way, though more complicated conditions may be encountered in different areas. The detailed discussions in individual cases are not attempted in the present note, because reliable data, ocean-

(2) Time change in vorticity produced in this manner, however, would be still negligibly small compared with each of horizontal divergence and wind stress curl. Although the secondary variable currents will produce the corresponding additional changes in vertical velocity, this additional divergence is negligible in surface layer.

ographical as well as meteorological, are not readily available at hand. Only some general remarks may be made briefly. As seen from (6), upwelling on the left and sinking on the right of the stream looking in direction of flow will strengthen the main stream. Such conditions will be favored when the atmospheric high pressure is on the right, and the low on the left. The magnitude of canges in current speed is measured by $\frac{\partial^2 \tau_y}{\partial x^2}$, that is, roughly by the relative intensity of the high and low, and also by the distance between them. Hence considerable fluctuations in currents may occur in the boundary regions between atmospheric high and low, where, in addition to the above effects, the direction of the winds roughly coincides with the current and a slight change in position or intensity of the atmospheric system might cause considerable variations in the oceanic currents. Now, it might be supposed that annual or yearly variations in the location and intensity of the large-scale atmospheric pressure system would cause the large-scale variations in the oceanic circulation over the wide areas of the oceans. For example,[3] nearly simultaneous changes in the oceanographic conditions on both sides of the North Pacific, off Japan and off California, might be caused by yearly changes in the position or intensity of the North Pacific High. These are as yet only speculative and the future works in cooperation between the western and the eastern portions of the Pacific would reveal the features.

In some relation to this, it might be worth mentioning to remark on the variations in the currents which have been reported in varions occasions of the passage of storms, or in asoociation with some atmospheric pressure field of shorter duration. It appears that the Gulf Stream and Kuroshio are intensified in the cases when the high pressure is offshore and the low is in coastal side of the Stream. It has also been observed that the northward current is intensified on the right hand side of the course of the storm. The variations associated with passage of storms or meteorological disturbances of duration of a week or shorter are somewhat different in their behaviors from those discussed in this paper. Mass redistributions may be done only by part before the atmospheric disturbance is removed. They should also be of practical importance on fishery problems and should require further studies by specifically planned observations.

In concluding the present discussions, it might be mentioned that some of the characteristic changes in California Current System could be predicted on the basis of oceanographic data, say, collected one month earlier.

Conclusion and Discussions

It should be too early to draw any conclusion at to seasonal fluctuations in the oceanic circulation on the basis of more or less hypothetical theory. What we need now indeed should, as every one does realize, be the more and better observations.

The outstanding features in seasonal changes in surface current system off the California coasts which have been revealed by the observations during years appear to suggest that the essential mechanism might be rather simple. The theory developed in the present paper seems to account for the main features of the variations, not only qualitatively but also quantitatively. For such system, dynamical equations and vorticity equations would merely represent the quasistationary equilibrium between the acting forces. Of course, they cannot alone tell how the currents change in response to wind fluctu ations. The changes could only be predicted by the prediction equation like (6) which is derived under certain *assumptions* concerning the changes in mass distributions in response to the changes in wind conditions. The results from the monthly cruises in the California waters appear to confirm that the assumptions may approximately be justified and that the prediction equation may be valid to a first approxi-

(3) Similar idea has been presented by Paul L. Horrer, the Scripps Institution of Oceanography, at the California Sardine Conference, July, 1953, Santa Maria, Calif.

mation. Its validity might fail in the other regions of the oceans, where different mechanism may prevail and data may not readily be available.

Achnowledments

The writer is indebted deeply to Paul. L. Horrer and Joseph L. Reid, who gave him, during his stay at La Jolla, many valuable suggestions and informations of the data in the California current regions.

References

Jacobs, W. C., 1939. Sea level departures on the California coast as related to the dynamics of the atmcsphere over the North Pacific Ocean. J. Mar. Res. **2** (3) : 181-194.

Munk, W. H., J. G. Pattullo, R. R. Revelle, and Elizabeth Strong, 1954. The anual change in sea level. (In press)

Sverdrup, H. U., and W. E. Allen, 1939. Distribution of diatoms in relation to the character of water masses and currents off Southern California in 1938. J. Mar. Res. **2** (2) : 131-144.

Sverdrup, H. U., and R. H. Fleming, 1941. The waters off the coast of Southern California, March to July, 1937. Scripps Inst. Oceanogr., Bull. **4** (10) : 261-378.

Yoshida, Kozo, and Han-Lee Mao, 1955. A new measurement of upwelling with examples in the California current region. (In press)

Yoshida, Kozo, 1955a. *Coastal upwelling* off the California coast. (In press)

1955b. A discussion on responses of the oceanic circulation in response tc a fluctuating wind stress. (unpublished)

1955c. Interaction between the mixed layer and the lower layer. (uupublished)

A Note on the Compressional Disturbances Accompanying the Tsunami*

By

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Abstract: As the natural case, when a broad bottom deformation superimposed by the secondary unevennesses or by the rather narrow mainly deformed part happens, the check whether the energy of the compressional disturbances may concentrate or may not, is required. Our check proves that no energy concentration may occur.

In Part II of the referred papers, we treated the wave form due to the compressional disturbances when the radii of the deformed areas are much larger (or smaller) than the water depths. In this paper, we will treat the case in which the radius of the deformed area and the water depth are of the same orders.

The ratio of them is, therefore, now

$$\frac{H}{a}=h\approx 1.$$

Then using the ordinary t and the asymptotic expansion of J_0, (19) (in Part I of the referred papers) can be written as

$$\frac{\pi a^2}{V}\xi_0(t)$$

$$=\sqrt{\frac{2}{\pi}}\frac{z_0}{h^3}\int_0^1 sds\int_0^\pi d\omega\int_0^\infty\sum_{n=1}^{\infty}\frac{(-1)^n(2n-1)\sin\left\{\sqrt{m^2+b_n^2}-\delta\frac{(m^2-b_n^2)}{\sqrt{m^2+b_n^2}}\right\}\frac{ht}{z_0}\cos\left(m\Omega-\frac{\pi}{4}\right)\sqrt{m}}{\sqrt{\Omega}(m^2+b_n^2)^3}\,dm$$

$$=\frac{z_0}{\sqrt{2\pi}\,h^3}\Re\,\frac{e^{-\pi/4\cdot i}}{i}\int_0^1 sds\int_0^\pi d\omega\,[I_1-I_2] \tag{54}$$

here
$$I_{1,2}=\int_0^\infty\sum_{n=1}^{\infty}\frac{\sqrt{m}\,(-1)^n(2n-1)}{\sqrt{\Omega}(m^2+b_n^2)^3}\exp\left(i\Omega\left\{m\pm K\sqrt{m^2+b_n^2}\mp K\delta\frac{(m^2-b_n^2)}{\sqrt{m^2+b_n^2}}\right\}\right)dm$$

$$K=\frac{Ut}{a\Omega} \tag{55}$$

Let us put
$$f(m)=m\pm K\sqrt{m^2+b_n^2}\mp K\delta\frac{(m^2-b_n^2)}{\sqrt{m^2+b_n^2}} \tag{56}$$

and we will differentiate it as follows,

$$f'(m)=\frac{1}{\sqrt{m^2+b_n^2}}\left[\sqrt{m^2+b_n^2}\pm Km\left\{1-2\delta+\delta\frac{(m^2-b_n^2)}{(m^2+b_n^2)}\right\}\right] \tag{57}$$

As $\left|\delta\frac{(m^2-b_n^2)}{(m^2+b_n^2)}\right|<\delta$, then $Km\left\{1-2\delta+\delta\frac{(m^2-b_n^2)}{(m^2+b_n^2)}\right\}>0$.

I_1 has, therefore, no stationary phase, and I_2 has one. We will represent the latter by m_s.

Then
$$m_s=m_{s0}+m_{s1}\delta+\cdots\cdots,\quad m_{s0}=\frac{b_n}{\sqrt{K^2-1}},\quad m_{s1}=\frac{b_n(3K^2-2)}{(K^2-1)^{3/2}} \tag{58}$$

If we assume the order of Ω as follows

$$\delta^2\ll 1/\Omega\ll\delta \tag{59}$$

* Received Aug. 3, 1955.

** Tokyo University of Fisheries.

then $$i\Omega f(m_s+\varepsilon)=i\Omega\{f(m_{s0})+\delta m_{s1}f'(m_{s0})\}+\frac{i\Omega\varepsilon^2}{2}\{f''(m_{s0})+\delta m_{s1}f'''(m_{s0})\}+\cdots\cdots \quad (60).$$

Omitting the small modifications on the amplitude, we get

$$\frac{\pi a^2}{V}\xi_0=\frac{z_0}{h^3}\Re\int_0^1 sds\int_0^\pi d\omega\sum_{n=1}^{\infty}\frac{(-1)^n(2n-1)\sqrt{m_{s0}}}{\Omega\sqrt{K}(m_{s0}^2+b_n^2)^{3/4}b_n}\exp\left(i\Omega\{f(m_{s0})+\delta m_{s1}f'(m_{s0})\}\right) \quad (61).$$

And considering that $f'(m_{s0})\sim O(\delta)$, the following relation is obtained

$$\frac{\pi a^2}{V}\xi_0=\frac{2z_0{}^3\sqrt{(Ut)^2-r^2}}{\pi h^4t^2a}\int_0^1 sds\int_0^\pi d\omega\sum_{n=1}^{\infty}\frac{(-1)^n}{b_n}\cos\Omega f(m_{s0}) \quad (62).$$

Here, since

$$\Omega f(m_{s0})=-\left\{\Omega b_n\sqrt{K^2-1}-\frac{\Omega b_n\delta(2-K^2)}{\sqrt{K^2-1}}\right\}$$

$$\fallingdotseq -b_n\left\{\frac{\sqrt{U^2t^2-r^2}}{a}+\frac{\delta(U^2t^2-2r^2)}{a\sqrt{U^2t^2-r^2}}+\frac{rs\cos\omega}{\sqrt{U^2t^2-r^2}}\right\},$$

$$\therefore\quad \frac{\pi a^2}{V}\xi_0=\frac{2z_0{}^2(1-\vartheta^2)}{\pi t\vartheta}\int_0^\pi d\omega\Big[-\frac{\cos\omega}{h}F\Big(\frac{\pi r}{2H}\sqrt{\frac{1}{\vartheta^2}-1}+\frac{\pi r}{2H}\delta\frac{(1/\vartheta^2-2)}{\sqrt{1/\vartheta^2-1}}+\frac{\pi\cos\omega}{2h\sqrt{1/\vartheta^2-1}}\Big)$$

$$+\frac{2}{\pi}\sqrt{\frac{1}{\vartheta^2}-1}\Big\{H\Big(\frac{\pi r}{2H}\sqrt{\frac{1}{\vartheta^2}-1}+\frac{\pi r}{2H}\delta\frac{(1/\vartheta^2-2)}{\sqrt{1/\vartheta^2-1}}+\frac{\pi\cos\omega}{2h\sqrt{1/\vartheta^2-1}}\Big)$$

$$-H\Big(\frac{\pi r}{2H}\sqrt{\frac{1}{\vartheta^2}-1}+\frac{\pi r}{2H}\delta\frac{(1/\vartheta^2-2)}{\sqrt{1/\vartheta^2-1}}\Big)\Big\}\Big]\Big/\cos^2\omega \quad (63).$$

Now, the elevation of the sea surface is given by

$$\zeta_0=\frac{1}{T}[\xi_0(t)-\xi_0(t-T)].$$

We can know from the above analysis, that the special feature that much energy can concentrate at any areas, does not appear when the radius of the deformed bottom and the water depth are of the same orders. And this may become the foundation upon which we can suppose that when a broad deformation superimposed by the secondary unevennesses or by the rather narrow mainly deformed part occurs, energy concentrations at any areas are not expected.

In this treaty, we adopted the notations used in the referred papers. Finally, the author's sincere thanks are due to the people of Geophysical Institute and Professor R. Takahashi of the Earthquake Research Institute, Tokyo University, who kindly discussed with him.

References

Miyoshi, H. (1954): "Generation of the Tsunami in Compressible Water (Part I)", Journ. Oceanogr. Soc. Jap. **10** (1), pp. 1~9.

Miyoshi, H. (1955): "Generation of the Tsunami in Compressible Water (Part II)", Rec. Oceanogr. Wor. Jap. **2** (1), pp. 49~56.

Hydrographic Conditions of the North-Western Pacific. Part 1*.

On the Temperature Change in the Upper Layer in Summer

By

Nobuo WATANABE**

Abstract: On the basis of the data obtained during the oceanographic observations carried out in the adjacent sea to the western Aleutian Islands in the early summer of 1953, the author has studied the mechanisms of temperature change in the surface layer, considering the heat exchange between the sea and the atmosphere and the heat transportation due to water movements and mixing. As the results of it, the following conclusions have been deduced.

(1) As the factors contributing to the variation of the surface temperature, the heat exchange between sea water and atmosphere above the sea surface, eddy conductivity, advection and vertical water movement should be taken into consideration.

(2) Although the magnitudes of advection, vertical eddy and vertical motion of water masses did not distribute always uniformly among different stations in the area under study, but the magnitudes of heat transportation owing to these factors and heat radiation received at the sea surface were frequently of the same order.

Further, sea fog formed within the area under study has been considered as an important factor for the appearence of oceanographic conditions in summer, contributing the transportation of heat to the sea surface.

1. Introduction

In the vicinity of the salmon fishing ground located in the adjacent sea to the western Aleutian Islands, surface water temperature in summer has been increased within a range of 1°C or so only in a period of a decade, accompanying rapid variations of oceanographic conditions (Watanabe, 1952). Especially, changes in temperature distributions should be noted as one of the factors relating to the establishment of fishing grounds, since locations of the fishing ground have just been transferred, accompanying with the changes in temperature distributions.

In the present paper, the author has studied the mechanisms of temperature changes in the upper layer in the area covering over the south-western region of Attu Island, using the data which were obtained during the oceanographical observations carried out aboard R. V. "Tenyō-maru" from June 18 to July 12, 1953 (Tokai Regional Fisheries Research Laboratory). In addition to this, he has also glanced at the transference in locations of the salmon fishing ground with the changes in temperature distributions.

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2. Basic Theories

For the purpose of considering the heat exchange in the upper layer, we may schematically assume a model of a water column as shown in Fig. 1, after Sverdrup (1946) and Armstrong (1947). Taking x-axis from the west to the east, y-axis from the south to the north and z-axis from the surface to the bottom, a relationship concerning the inflow and outflow of heat into and out of the assumed water column may be obtained as expressed in the following equation, when various kinds of heat—namely, heat absorbed or radiated at the sea surface into or out of the column, heat transported by advection through the sides of the column, heat lossed from the bottom surface of the column and heat remaining in the column, causing tem-

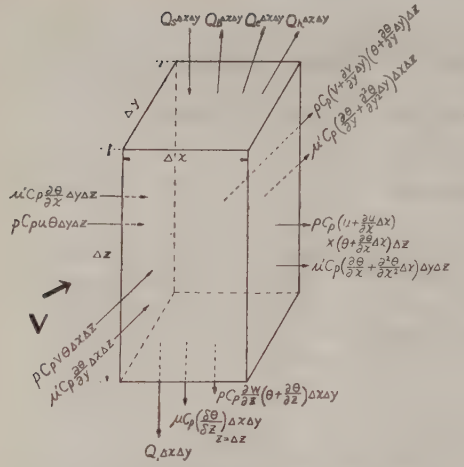


Fig. 1. Schematic model of heat exchange at the surface layer of the ocean.

perature change of sea water—are taken into consideration.

$$\int_0^{\Delta t}\rho c_p\left(\frac{\partial\theta}{\partial t}\right)\Delta x\Delta y\Delta z dt$$
$$=\int_0^{\Delta t}\left\{(Q-Q')-\mu c_p\left(\frac{\partial\theta}{\partial z}\right)_{z=\Delta z}\right\}\Delta x\Delta y dt$$
$$-\int_0^{\Delta t}\mu' c_p\left\{\left(\frac{\partial^2\theta}{\partial x^2}\right)_{x=\Delta x}\Delta x+\left(\frac{\partial^2\theta}{\partial y^2}\right)_{y=\Delta y}\Delta y\right\}\Delta z dt$$
$$-\int_0^{\Delta t}\rho c_p\left\{\left(u\frac{\partial\theta}{\partial x}\Delta x+v\frac{\partial\theta}{\partial y}\Delta y\right)\Delta z\right.$$
$$\left.+\left(\theta\frac{\partial u}{\partial x}\Delta x+\theta\frac{\partial v}{\partial y}\Delta y\right)\Delta z\right.$$
$$\left.+\left(\theta\frac{\partial w}{\partial z}\Delta z\right)\Delta x\Delta y\right\}dt$$

where the meanings of each notation are as follows:

Q ; heat absorbed to sea water through the sea surface ($Q=Q_s-Q_b-Q_e-Q_h$) (g.-cal./cm².day),

Q_s ; heat radiated from the sun and sky (g.-/cal./cm².day),

Q_b ; heat radiated from the sea surface to the atmosphere (g.-cal./cm².day),

Q_e ; heat lossed from the sea surface by evaporation (g.-cal./cm².day),

Q_h ; heat lossed by the process of convection in the atmosphere (g.-cal./cm².day),

Q' ; heat transferred through the bottom surface of the water column (g.-cal./cm².day),

ρ ; density of sea water,

c_p ; specific heat of sea water at constant pressure,

θ ; temperature of sea water,

μ ; vertical eddy conductivity of heat in sea water,

μ' ; lateral eddy conductivity of heat in sea water,

Δt ; time interval in which the temperature change has been caused,

$\int_0^{\Delta t}\rho c_p\left(\frac{\partial\theta}{\partial t}\right)\Delta x\Delta y\Delta z dt$; heat required for the temperature change of sea water in the column within the period of Δt,

$\int_0^{\Delta t}(Q-Q')\Delta x\Delta y dt$; difference between radiant heat received at the surface and heat transferred out of the column through the bottom surface, that is, the ultimate radiant heat remaining in the water column,

$\int_0^{\Delta t}\mu c_p\left(\frac{\partial\theta}{\partial z}\right)_{z=\Delta z}\Delta x\Delta y dt$; heat lossed through the bottom surface by vertical eddy conductivity,

$\int_0^{\Delta t}\mu' c_p\left\{\left(\frac{\partial^2\theta}{\partial x^2}\right)_{x=\Delta x}\Delta x+\left(\frac{\partial^2\theta}{\partial y^2}\right)_{y=\Delta y}\Delta y\right\}\Delta z dt$; heat lossed through the sides of the column by lateral eddy conductivity,

$\int_0^{\Delta t}\rho c_p\left(u\frac{\partial\theta}{\partial x}\Delta x+v\frac{\partial\theta}{\partial y}\Delta y\right)\Delta z dt$; heat transported out of the column by lateral advection,

$\int_0^{\Delta t}\rho c_p\left(\theta\frac{\partial u}{\partial x}\Delta x+\theta\frac{\partial v}{\partial y}\Delta y\right)\Delta z dt$; heat transported out of the column by the divergence of water movement,

$\int_0^{\Delta t}\rho c_p\left(\theta\frac{\partial w}{\partial z}\Delta z\right)\Delta x\Delta y dt$; heat transported out of the column through the bottom surface by the vertical motion of water.

Now, if we assume $\Delta x=\Delta y=1$, the equation mentioned aboved above is reduced in the from:

$$\int_0^{\Delta t}\rho c_p\left(\frac{\partial\theta}{\partial t}\right)\Delta z dt$$
$$=\int_0^{\Delta t}\left\{(Q-Q')-\mu c_p\left(\frac{\partial\theta}{\partial z}\right)_{z=\Delta z}\right\}dt$$
$$-\int_0^{\Delta t}\mu' c_p\left\{\left(\frac{\partial^2\theta}{\partial x^2}\right)+\left(\frac{\partial^2\theta}{\partial y^2}\right)\right\}\Delta z dt$$
$$-\int_0^{\Delta t}\rho c_p(\boldsymbol{v}\cdot\nabla\theta)\Delta z dt$$
$$-\int_0^{\Delta t}\rho c_p\left\{\theta\left(\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}+\frac{\partial w}{\partial z}\right)\Delta z\right\}dt\cdots\cdots \quad (1).$$

Since density of sea water may be considered as constant during the time interval of Δt and the condition of continuity, $\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}+\frac{\partial w}{\partial z}=0$, may be satisfied in this case, the Eq. (1) can be written

$$\int_0^{\Delta t}\rho c_p\left(\frac{\partial\theta}{\partial t}\right)\Delta z dt = \int_0^{\Delta t}\left\{(Q-Q')-\mu c_p\left(\frac{\partial\theta}{\partial z}\right)_{z=\Delta z}\right\}dt - \int_0^{\Delta t}\mu' c_p\left\{\left(\frac{\partial^2\theta}{\partial x^2}\right)+\left(\frac{\partial^2\theta}{\partial y^2}\right)\right\}\Delta z dt - \int_0^{\Delta t}\rho c_p(\boldsymbol{v}\cdot\nabla\theta)\Delta z dt \qquad (2).$$

In the Eq. (2), the second term in the right side of the equation may be ignored since the values of $\frac{\partial^2\theta}{\partial x^2}$ and $\frac{\partial^2\theta}{\partial y^2}$ are very small generally,* even if we assume that $\frac{\mu'}{\mu}\fallingdotseq 10^8$. Then

$$\int_0^{\Delta t}\rho c_p\left(\frac{\partial\theta}{\partial t}\right)\Delta z dt = \int_0^{\Delta t}\left\{(Q-Q')-\mu c_p\left(\frac{\partial\theta}{\partial z}\right)_{z=\Delta z}\right\}dt - \int_0^{\Delta t}\rho c_p(\boldsymbol{v}\cdot\nabla\theta)\Delta z dt \qquad (2)'.$$

In the Eq. (2′), we may consider $c_p=$const. and $\Delta z=10^3$ cm., since we have considered of the upper layer whose thickness is 10 m. Further, since

$\int_0^{\Delta t}\left(\frac{\partial\theta}{\partial t}\right)dt=\Delta\theta$; (temperature change in the interval of Δt),

$\int_0^{\Delta t}(Q-Q')\,dt=(\overline{Q}-\overline{Q}')\Delta t$; $\overline{Q}$ and $\overline{Q}'$ are the mean values in the interval of Δt,

$$\int_0^{\Delta t}\mu c_p\left(\frac{\partial\theta}{\partial z}\right)_{z=\Delta z}dt=c_p\int_0^{\Delta t}\mu\left(\frac{\partial\theta}{\partial z}\right)_{z=\Delta z}dt = c_p\cdot\bar{\mu}\cdot\frac{1}{2}\left\{\left(\frac{\partial\theta}{\partial z}\right)_1+\left(\frac{\partial\theta}{\partial z}\right)_2\right\}\cdot\Delta t;$$ (heat transported by vertical eddy conductivity),

and $$\int_0^{\Delta t}\rho c_p(\boldsymbol{v}\cdot\nabla\theta)\Delta z dt = \frac{1}{2}\rho c_p\{(\boldsymbol{v}\cdot\nabla\theta)_1+(\boldsymbol{v}\cdot\nabla\theta)_2\}\Delta z\Delta t;$$

(heat transported by lateral advection), the following equations will be obtained, introducing these terms into the Eq. (2′).

$$\mu=\frac{1}{\frac{1}{2}\left\{\left(\frac{\partial\theta}{\partial z}\right)_1+\left(\frac{\partial\theta}{\partial z}\right)_2\right\}}\cdot\left\{\left(\frac{Q-Q'}{c_p}\right)-10^3\rho\left(\frac{\Delta\theta}{\Delta t}\right)-\rho\cdot\frac{1}{2}\{(\boldsymbol{v}\cdot\nabla\theta)_1+(\boldsymbol{v}\cdot\nabla\theta)_2\}10^3\right\} \qquad (3).$$

$$\left(\frac{\Delta\theta}{\Delta t}\right)=\frac{10^{-3}}{\rho}\left\{\left(\frac{\overline{Q}-\overline{Q}'}{c_p}\right)-\bar{\mu}\,\frac{1}{2}\left(\left(\frac{\partial\theta}{\partial z}\right)_1+\left(\frac{\partial\theta}{\partial z}\right)_2\right)\right\}-10^3\cdot\frac{1}{2}\{(\boldsymbol{v}\cdot\nabla\theta)_1+(\boldsymbol{v}\cdot\nabla\theta)_2\} \qquad (4).$$

3. Computations concerning the inflow and outflow of radiant heat through the sea surface

As for the values of $\overline{Q}-\overline{Q}'$ in the preceding section, the present author has made an estimation after Sverdrup (1946), Jacobs (1942) and Miyazaki (1952), evaluating several terms which will be stated below.

(i) Q_s (heat radiated from the sun and sky) is expressed by Mosby and Schmidt in the form:

$$Q_s=k\left(1-0.71\,\frac{C}{100}\right)\bar{h}\,(1-r)\times 1440,$$

(g.-cal./cm².day),

where, $k=0.025$ (a coefficient concerning transparency of the atmosphere given by Mosby, which was evaluated for lat. 50°N),

$h=23°$ (mean altitude of the sun computed for the station situated on lat. 51°N and long. 167°E in the last decade of June, 1953),

$C=$cloudiness % (values observed aboard "Tenyō-Maru"),

and $r=11\%$ (reflection rate at the sea surface given by Schmidt).

(ii) Q_b (heat lossed by the radiation from the sea surface is given by

* $\frac{\partial^2\theta}{\partial x^2}$ or $\frac{\partial^2\theta}{\partial y^2}:\frac{\partial^2\theta}{\partial z^2}\fallingdotseq 1:10^{10}$; (Calculated values by the data during the oceanographic observation by R. V. "Tenyō-maru" mentioned above). However in the case of water boundary the circumstances may be somewhat different one.

$Q_b = 0.94\,\sigma T_w^4 (1-0.44-0.08\sqrt{e_a})\,(1-0.0083\,C) \times 1440$, (g.-cal./cm².day),

where, $\sigma = 82\times10^{-12}$ (Stefan's constant),
$T_w = 273+\theta_w$°C (absolute temperature of sea surface)
and e_a = tension of water vapour in the atmosphere in mb. (values observed aboard "Tenyō-maru").

(iii) R (Bowen Ratio)

$$R = 0.66\,\frac{P}{1000}\cdot\frac{\theta_w-\theta_a}{e_w-e_a} = \frac{Q_h}{Q_e},$$

where, P = atmospheric pressure in mb. (values observed aboard "Tenyō-maru"),
θ_w = temperature of sea water in °C (observed values aboard "Tenyō-maru"),
θ_a = temperature of the air in °C (values observed aboard "Tenyō-maru"),
$e_w = e_{\theta_w\text{-max}}$ mb. (maximum tension of water vapour corresponding to the water temperature),
and e_a = tension of water vapour in the air, mb. as mentioned above.

(iv) E (evaporation from the sea surface, mm./cm²./day)

$E = 0.143\,(e_w-e_a)\,W_a$ ……Jacobs' equation*

where W_a = wind velocity in meters per second (values observed aboard "Tenyō-maru").

(v) Q_e (heat lossed by the process of evaporation, g.-cal./cm².day)

$$Q_e = L_t \cdot E \cdot 10^{-1},$$

where L_t = latent heat of evaporation = $596 - 0.52\,\theta_w$.

(vi) Q_h (heat lossed by the process of convection g.-cal./cm².day)

$$Q_h = R \cdot Q_e.$$

(vii) Q (heat absorbed through the sea surface into the water column assumed, g.-cal./cm².day)

$$Q = Q_s - Q_b - Q_e - Q_h.$$

(viii) Q' (radiant heat transmitted downward through the lower surface of the upper layer, g.-cal./cm².day).

Radiant heat received at the sea surface will be absorbed by molecules and ions in sea water when it enters into the liquid medium even if no disturbance on the surface exists. Hence, Q', heat transmitted through a layer of Δz, is given by the equation:

$$Q' = \alpha_{\Delta z} \cdot Q,$$

where $\alpha_{\Delta z=10\,m}$ has a value of 0.161 in transparent oceanic water (Sverdrup, 1956).

The author carried out computations to evaluate each term mentioned above, using data of observations made aboard R.V. "Tenyō-maru" in 1953, results of which are shown in Table 1. The table shows characteristic features concerning the inflow and outflow of heat at the sea surface in the north-western Pacific in summer months. In the table, the whole period during which the observations carried out were separated into two periods, depending on meteorological conditions. In short, the first or second period corresphnds to early or mid summer in the area under study, respectively. During the first period the wind velocity and cloudiness were greater, and air temperature

* Jacobs (1942) applied the equation to the region covering lat. 40–45°N and long. 160°E–140°W.

Table 1. Radiant Heat at the Surface of the North-Western Pacific (Last decade of June~1st decade of July, 1953)

Items / Period	Q_s g.-cal./cm²/day	Q_b g.-cal./cm²/day	Q_e g.-cal./cm²/day	Q_{ie} g.-cal./cm²/day	Q g.-cal./cm²/day	Q' g.-cal./cm²/day	R	E mm./cm²/day
1st (12, June~1, July)	214.56	37.30	−19.93	19.90	177.29	28.37	−0.9986	−0.34
2nd (7~12, July)	344.16	83.70	5.33	9.88	245.25	39.24	+1.850	+0.09
Whole Period	266.03	53.54	−10.97	17.03	211.27	33.80	−1.554	−0.19

colder, more foggy, than these observed during the second period, while the air temperature became higher in the second period with the decreased effect of these meteorological factors. It may be clear that the meteorological factors affect the variations in various kinds of heat radiation. Namely, Q_s may increase with the decrease of C, sudden increase in Q_e may be caused chiefly by the activated process of evaporation and so on. And it should be also pointed out that the inversion of sign took place as to Bowen Ratio between the two periods. During the first period, heat lossed by convection and heat lossed by evaporation appear having opposite sign each other. Here, Q_e having negative sign indicates that much of heat of condensation enters into the sea water from the atmosphere. That is sea fog which appears in early summer may be considered as an important factor in supplying heat to the sea surface. On the contrary, Q_e has positive values during the second period, indicating the loss of heat by evaporation.

As for Q_h, the amount computed for the first period are much greater than that for the second, although outflow of heat was taking place in both periods. Thus, the ultimately total heat Q supplied to the sea surface in the second period had increased about the half of that in the first period.

Throughout the two periods, the author studied the variations of water temperature in the upper layer, assuming the mean rate of the heat 211.27 g.-cal./cm^2./day was supplied to the sea surface continuously during the time interval of Δt. The author also assumed that Q', which transmitted through a layer of 10 m.-deep, was 33.80 g.-cal./cm^2./day, having no effect on the temperature change in the surface layer.

4. An estimation on the coefficient of vertical eddy conductivity

In order to estimate the values of μ, the quantities required for the computation using the Eq. (3) have been obtained from distribution diagrams concerning water temperature and current* (Figs. 2 and 3), except for the values of $Q-Q'$ which were obtained in the preceding section. The computations were made on 16 stations located the area under study (Fig. 4). The procedures for the evaluation of each term are as follows;

$\frac{1}{2}\left\{\left(\frac{\partial\theta}{\partial z}\right)_1+\left(\frac{\partial\theta}{\partial z}\right)_2\right\}$: the vertical gradients of water temperature during the first and second period were averaged for each station.

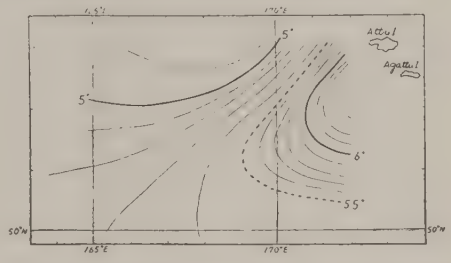


Fig. 2 a. Surface temperature °C (1st. period)

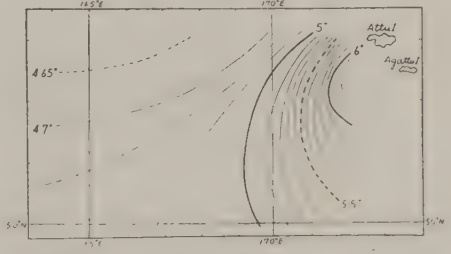


Fig. 2 b. Temperature at 10 m. depth °C (1st.i perod)

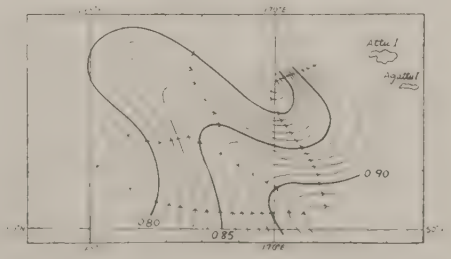


Fig. 2 c. Current by ΔD [based 800 m. depth] (1st. period)

* In case of the dynamical computation of currents the effect of the wind-drift can be neglected, since the wind velocities were almost less than 5 m./sec. during these periods in question (cf. Appendix :Tab.).

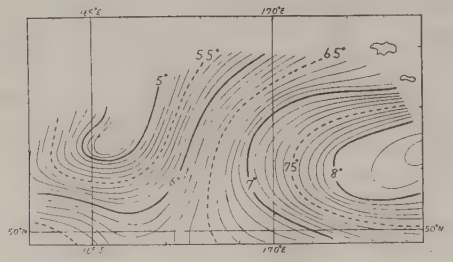


Fig. 3a. Surface temperature °C (2nd. period)

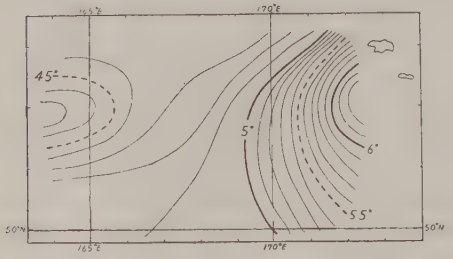


Fig. 3b. Temperature of 10 m. depth °C (2nd. period)

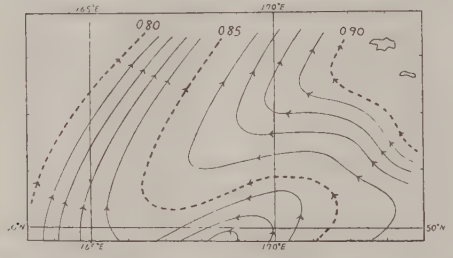


Fig. 3c. Current by ΔD [based 800. m depth] (2nd. period)

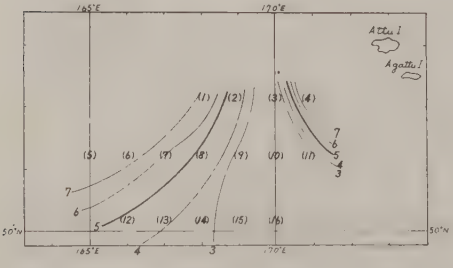


Fig. 4. Distribution of μ 0-10 m. (C.G.S. unit) and location.

ρ: density of sea water was considered as constant throughout the periods within the whole area, applying a value of 1.026.

Cp: specific heat of sea water at constant pressure was regarded as constant, applying a value of 0.939.

$\overline{Q}-\overline{Q'}$: heat remaining in a water layer between the surface and 10 m. was assumed to have a value of 211.27-33.80 (g.-cal./cm./day) obtained in the preceding section.

$\Delta\theta$: change in temperature took place during a period of 13.5 days (°C/day) for each station.

$\frac{1}{2}\{(\boldsymbol{v}\cdot\nabla\theta)_1+(\boldsymbol{v}\cdot\nabla\theta)_2\}$: scalar products of horizontal gradients concerning current velocity and water temperature observed at each station were averaged throughout the first and second periods.

$\bar{\mu}$: coefficients of vertical eddy conductivity (g.-cal./cm^2./day) for each station throughout the periods have been computed using the each term mentioned above, results of which are given in Table 2. Spacial distribution of these values are shown in Fig. 4. On the basis of these tables and figures, it may be concluded that the values estimated differ from each other for each station and can hardly be regarded as a constant. Namely, the values estimated for the southeastern part of the area are smaller than those of the vicinity of Attu or Commandor Islands.

Magnitudes of these values are ranging from 18×10^4 to 63×10^4 g.-cal./cm^2./day (2.1-7.3 in C.G.S. units), being in the comparable order with the values hitherto estimated.*

5. Relationships among the various kinds of heat contributing the changes of water temperature in the upper laper

In order to examine the mechanisms of contributing interactions between heat received at the sea surface and heat transported by advection of water masses or eddy conductivity to the changes of temperature in the upper layer, the author computed the each term in the Eq. (4), assuming that thermal equilibrium had been established at each station for which the values of μ were estimated. The results

* Jacobsen (1913); All current, 0-15 m., 1.9-3.8 in. C.G.S. units (in Danish water, very great stability), (cf. Sverdrup "The Oceans").

Table 2. Heat Transportation of several factors and surface temperature change at the surface layer in the North-Western Pacific, early summer, 1953

St. No.	μ 0–10m. (C.G.S.)	Heat g.-cal./cm².day							Contributions to the temperature change (°C/Δt)						Ratio of the each Heat for the receipt at the surface (%)						
		Receipt at the surface	Temperature Change	Lateral Transport	Vertical Transport	the items of Vertical Transp.			Receipt at the surface	Lateral Transport	Vertical Transport	the items of Vertical Transp.			Temperature increase	Receipt at the surface	Lateral Transport	Vertical Transport	the items of Vertical Transp.			Temperature increase
						Eddy	Current	Transmission				Eddy	Current	Transmission					Eddy	Current	Transmission	
1	6.9	211.3	63.6	+10.6	+128·1	+134.1	−30.8	33.8	+3.0	−0.2	−1.9	−1.9	+0.4	−0.5	+0.9	100	5	65	64	−15	16	30
2	4.1	〃	84.8	+57.8	+68.7	+86.0	−51.1	〃	〃	−0.8	−1.0	−1.2	+0.7	〃	+1.2	〃	27	33	41	−24	〃	40
3	2.3	〃	106.9	+75.1	+29.2	+55.1	−59.7	〃	〃	−1.1	−0.4	−0.8	+0.8	〃	+1.5	〃	36	14	26	−28	〃	51
4	7.3	〃	80.0	+61.7	+69.6	+154.3	−118.5	〃	〃	−0.9	−1.0	−2.2	+1.7	〃	+1.1	〃	29	33	73	−56	〃	38
5	7.3	〃	6.7	−7.7	+212.3	+172.7	+5.8	〃	〃	+0.1	−3.0	−2.4	−0.1	〃	+0.1	〃	−4	100	82	3	〃	3
6	7.0	〃	12.5	−71.3	+270.0	+176·5	+59.7	〃	〃	+1.0	−3.8	−2.5	−0.8	〃	+0.2	〃	−34	128	83	28	〃	6
7	5.5	〃	43.4	+8.7	+159.2	+147.6	−22.2	〃	〃	−0.1	−2.2	−2.1	+0.3	〃	+0.6	〃	4	75	70	−11	〃	21
8	4.2	〃	67.4	−48.2	+192.0	+118.7	+39.5	〃	〃	+0.7	−2.7	−1.7	−0.6	〃	+0.9	〃	−23	91	56	19	〃	32
9	2.7	〃	101.2	133.0	−22.9	+80.1	−136.8	〃	〃	1.9	+0.3	−1.1	+1.9	〃	+1.4	〃	63	−11	38	−65	〃	48
10	2.1	〃	120.4	+130.1	−39.2	+58.0	−131.0	〃	〃	−1.8	+0.5	−0.8	+1.8	〃	+1.7	〃	62	−19	28	−62	〃	57
11	2.2	〃	135.8	+33.7	+41.7	+42.6	−34.7	〃	〃	−0.5	−0.6	−0.6	+0.5	〃	+1.9	〃	16	20	20	−17	〃	64
12	4.4	〃	59.7	−9.6	+161.2	+123.5	+3.9	〃	〃	+0.1	−2.3	−1.7	−0.1	〃	+0.8	〃	−5	76	58	2	〃	28
13	4.0	〃	62.6	+14.5	+134.2	+117.7	−17.3	〃	〃	−0.2	−1.9	−1.6	+0.2	〃	+0.9	〃	7	64	56	−8	〃	30
14	3.4	〃	78.0	−6.7	+140.0	+102.3	+3.9	〃	〃	+0.1	−2.0	−1.4	−0.1	〃	+1.1	〃	−3	66	48	2	〃	37
15	2.8	〃	90.6	−128.1	+248.8	+84.0	+131.0	〃	〃	+1.8	−3.5	−1.2	−1.8	〃	+1.3	〃	−61	118	40	62	〃	43
16	2.7	〃	98.3	−64.6	+177.6	+72.5	+71.3	〃	〃	+0.9	−2.5	−1.0	−1.0	〃	+1.4	〃	−31	84	34	34	〃	47

Table 3. Contribution of the Heat Exchange between the water surface and atmosphere to the surface water temperature changes in the North-Western Pacific.
(Last decade of June~1st decade of July, 1953)

Items	Q Received at the surface	Q_{vt} Vertical transmission	Q_{ht} Lateral Transport	Q_{ve} Transport by Vertical Eddy	Q_{vc} Transport by Vertical current	Q_c Contribution to the water temp. change
Heat g.-cal./cm².day	+211.3	−33.8	−128.1~+133.0	−42.6~−172.7	−136.8~+131.0	+6.7~+135.8
Ratio Q_t/Q %	+100	−16	−61~+63	−20~−82	−65~+62	+3~64

Remarks: The sign put on the numerals in the table "+" or "−" shows the contribution to temperature increase or decrease respectively.

of the computations were shown in Table 2. Here, according to the assumption:

$$\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}+\frac{\partial w}{\partial z}=0,$$

heat transported by lateral divergence was expressed as that by upwelling, since lateral divergence would occur in the region within which upwelling was taking place. On the contrary, heat transported by lateral convergence was expressed by giving negative sign to heat transported by vertical water movement, although heat transportation due to lateral convergence was regarded as that owing to lateral advection, since convergence would correspond to sinking of water masses. Thus, mechanisms of the changes in water temperature in the upper layer may be formulated as follows:

temperature changes = the one due to the heat received at the sea surface + the one due to the lateral transportation + the one due to the vertical transportation

vertical heat due to transportation = heat due to transportation by vertical eddy condition + heat due to transportation by vertical water movement + heat due to vertical transmission

As described above, the changes of water temperature due to radiant heat at the sea surface should not be overlooked in the upper layer of the north-western Pacific in summer months, together with other affecting factors. Since heat received at the sea surface may be regarded as uniform over a broard area, the heterogeneous distribution of water temperature in the area under study should be attributed to other factors. However, the governing factor is not always identical for each station, that is, at some station the effect of lateral advection is prevailing, at another station the effect of vertical transportation is a governing factor and so on. In case of the temperature changes due to vertical transportation of heat, the temperature change caused by upwelling or sinking of water masses should not be overlooked, although heat transported vertically by eddy conductivity is not uniform, depending on the mutual difference of the magnitudes of $\bar{\mu}$. Table 3 were constructed in order to obtain a general view on magnitudes of heat transportation by the various factors in the area under study.

For the purpose of considering the problem more precisely, the whole changes of water temperature in the upper layer have been separated into several parts, each of which has been including a single factor as a cause of temperature change. The separated temperature changes were illustrated in Fig. 5-8 for each single factor.

In the distribution diagram of temperature changes caused by lateral advection (Fig. 5), two regions of rising temperature (0.5-1.0°c/13.5 days) exist covering the north eastern and central parts of the area and off Attu Island, while positive temperature change is not recognized for the regions in the northern edge and the south-western corner of the area.

Fig. 7 shows the distribution of temper-

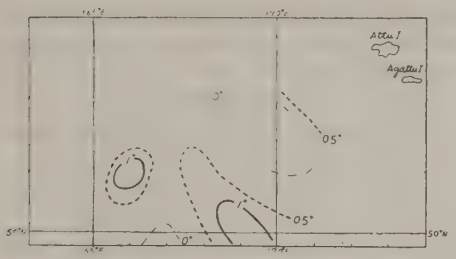


Fig. 5. Surface temperature changes °C by the lateral advection.

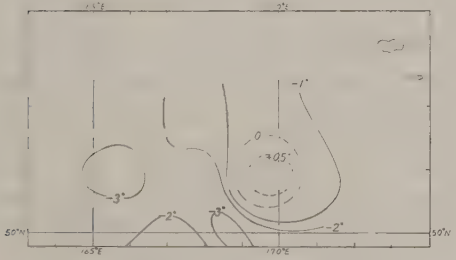


Fig. 6. Surface temperature changes °C by total vertical transport of heat.

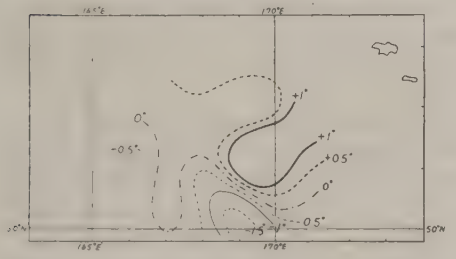


Fig. 7. Surface temperature changes °C by the vertical water movement.

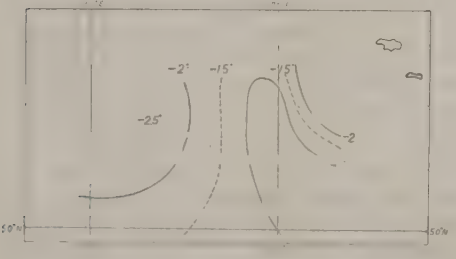


Fig. 8. Surface temperature changes °C by the vertical eddy motion.

ature change due to the vertical water movement. There are two characteristic regions of rising or falling temperature. As the vertical gradicuts of temperature $\left(\frac{\partial\theta}{\partial z}\right)$ are negative in the upper layer throughout the area, the region of rising temperature corresponds to that of sinking where the temperature increases 0.5°–1°C in a period of 13.5 days, and the region of falling temperature to that of upwelling where the temperature is remarkably decreased within a range from 0.5° to 1.5°C in the same period. Then, comparing Fig. 7 with Fig. 5, it should be noticed that the region of vigorous upwelling corresponds to that of increasing temperature due to lateral advection, while in the region of sinking no marked positive change of temperature due to lateral advection is recognized.

As to the variation of temperature only due to vertical eddy conductivity (Fig. 8), it should be pointed out that in the region within which the lateral advection and the upwelling are prevailling as mentioned above, the magnitudes of ultimate variation are small, while in the region within which the lateral advection and the sinking are weakened, they indicate markedly great, being a cause of falling temperature.

Fig. 6 is the synoptic chart denoting the variation of temperature caused by factors concerning variations along the vertical direction. The figure shows total variation brought by the vertical transportation of heat, including the facts that the greater the effect of lateral advection, the greater the variation.

Thus, the changes of water temperature in the upper layer throughout the periods should be determined according to the magnitudes of remaining heat in the layer, which may result from the processes of the inflow or outflow of heat into or out of the layer. In connection with this, it should be, as mentioned, an interesting fact that the salmon fishing grounds during the periods have had their centers mainly in regions in which upwelling was most prevailing (Fig. 9). These regions correspond to those in which the variation of water temperature due to vertical transportation of heat is prevailing, although the lateral motion of water is also remarkable.

Notwithstanding the considerable vari-

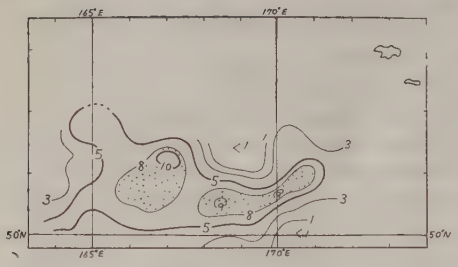


Fig. 9. Location of the salmon fishing ground during that last decade of June to the first decade of July in 1953. (Numerals show the number of catch per sheet of drift net).

ations in the lateral distribution of water temperature are noticed when the time elasped from the first period to the second, the transportations of water masses during these periods cannot be regarded as the ouly cause of the variation. Thus, it may be difficult to discuss the transportations of water masses simply by transference of iso-thermal lines.

Fig. 10 has been plotted by means of adding the temperature variation only

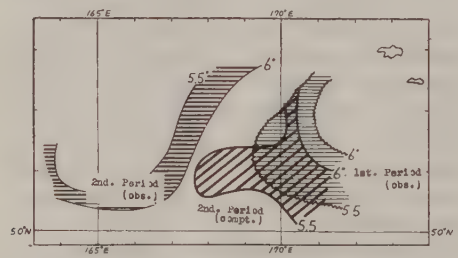


Fig. 10. Comparison of the transference of the zonal domain (5.5°–6.0°C) by computation of transport merely due to the advection with those of actual observation in the whole time interval Δt.

owing to currents to the distribution of temperature in the first period and comparing with the distribution in the second period. For convenience of comparison of the distributions with each other, the variations of the zonal domains enclosed by the iso-thermal lines of 5.5 and 6.0°C had been remarked. Although the zonal domain was transferred westwards in the couse of time, the distance of transference was much shorter than that estimated on the basis of the apparent difference between the temperature distributions in the first and second periods. On the whole, it may be impossible to discuss the transportations of water masses in the area under study on such a expedient basis even if the term "transportation" were understood as that caused simply by lateral advection, upwelling and sinking, because the radiant heat at the sea surface and vertical eddy conductivity affect greatly on the transference of iso-thermal lines.

6. Summary

The author studied the mecanisms of temperature changes in the upper layer of the adjacent sea to the western Aleutian Islands in the summer of 1953, using the theory on thermal equilibrium in the upper layer. As a result, the following conclusions have been derived:

(i) As factors contributing to the temperature changes in the upper layer, heat exchange between the sea and atmosphere, eddy conductivity, lateral advection and vertical water movement should be taken into consideration.

(ii) The estimated magnitudes of the contributing factors, such as lateral advection, vertical eddy and vertical water movement, to the temperature change differ greatly from each other for each station.

However, the heat transported by these factors can become comparable with the heat received at the sea surface.

(iii) The estimated magnitudes of temperature changes due to lateral mixing were negligibly small in the sea under investigation.

(iv) Sea fog formed in the area under study contributes to the supply of heat to the sea surface, seems to be one of important factors which increases the water temperature of surface layer in the process from the spring to the summer.

(v) The main fishing grounds of salmon have their centers in the region in which upwelling is taking plance remarkably.

(vi) It is difficult to discuss the transportation of water masses merely on the

basis of apparent transference of isothermal lines.

7. Acknowledgement

In conclusion the author wishes to acknowledge his indebtedness to Director I. Genshō, Tokai Regional Fisheries Research Laboratory, Prof. Dr. M. Uda, Tokyo Fisheries University, for their kindness in giving much valuable advice and in correcting the manuscript. Further he must express his gratitude to Dr. M. Migita, Mr. Hirano, Mr. R. Fukai and other members in the Oceanographical Section of T.R.F.R.L. who have supported him in many ways during his work.

References

Armstrong, J. C., 1947: The maintenance of instability in the surface waters of the ocean. Scripps Inst. of Oceanography-Contribution No. 313, 1947 (Reprinted from The Annales of the New York Academy of Science Vcl. XLVIII, Art. 8, Sept. 15, 1947).

Jacobs, 1942: On the energy exchange between sea and atmosphere. Jour. Mar. Res., 5, No. 1, 1942.

Miyazaki, M.: The Heat Budget of the Japan Sea. Bulletin of Hokkaido Regional Fisheries Research Laboratory, No. 4, March, 1952. H.R.F.R.L., Hokkaido, Japan (in Japanese).

Sverdrup, H.U. et al, 1946: "The oceans, their physics, chemistry and general biology" Prentice-Hall, INC., New York.

Tokai Regional Fisheries Research Laboratory: The observed data of the oceanographical investigation carried out by the R. V. "Tenyō-maru" in the North-Western Pacific during Summer in 1953, Tokai Regional Fisheries Research Laboratory, Tokyo, Japan (not yet issued).

Watanabe, N., 1952: A report on the oceanographical observations made in the salmon fishing ground located in the north-western Pacific from May to July, 1952. (in Japanes). Dec. 1952, Tokai Regioral Fisheries Research Laboratory, Tokyo, Japan.

Appendix: Table of Observational Data aboard the R. V. "Tenyo-maru" in the North-Western Pacific during eary summer, 1953

Period	St.	Date	Location		θ°C		S‰		θ_α °C	θ_α'°C *	C%	W_a m/s	P_{mb}
			Lat. L	Long. E	0m	10m	0m	10m					
		M d	° ′	° ′									
the 1st (18, June~1, July)	25	6-18	52-46	170-48	5.40	5.07	33.22	33.22	6.2	5.8	100	7.0	1017.2
	27	20	50-17	171-13	5.40	5.11	33.12	33.04	4.6	4.4	100	4.5	1019.9
	28	26	51-10	164-00	5.20	4.79	33.22	33.22	7.8	6.4	100	1.0	1015.3
	29	27	51-07	165-22	5.20	4.82	33.13	33.15	5.6	5.4	100	4.5	1018.9
	30	29	52-03	166-12	5.00	4.67	33.08	33.10	5.6	5.4	100	9.4	1013.2
	31	30	52-57	166-07	4.90	4.63	33.10	33.06	5.8	5.6	100	9.4	1012.6
	32	7- 1	52-32	168-38	4.90	4.63	33.04	33.10	6.0	5.8	100	9.4	1013.2
	33	1	52-22	170-10	5.40	5.12	33.22	33.22	6.2	6.0	100	4.5	1017.2
	34	1	52-22	171-55	6.40	6.24	33.01	33.01	6.0	5.8	100	4.5	1019.9
	Mean				5.32	5.01	33.13	33.12	6.0	5.9	100	5.0	1016.3

$\bar{\theta}_{w\cdot 0-10m}=5.16$°C $\quad \theta_{w\cdot 0}-\theta_{w\cdot 10}=0.31$°C $\quad \bar{\theta}_\alpha-\bar{\theta}'_\alpha=0.1$°C $\quad e_\alpha=0.12$ mb

$\bar{S}_{0-10}=33.13$‰ $\quad \sigma_t=26.21$ $\quad e_w=0.98\ (e_{\theta w\cdot \max})=8.65$ mb

Period	St.	Date	Location		θ°C		S‰		θ_α °C	θ_α'°C *	C%	W_a m/s	P_{mb}
			Lat. L	Long. E	0m	10m	0m	10m					
		M d	° ′	° ′									
the 2nd (5~12, July)	35	7- 5	52-21.5	173-28	6.70	6.86	33.19	33.17	6.8	6.2	unknown	2.0	1017.2
	36	8	52-17	173-24	7.50	7.01	33.10	33.12	9.2	8.4	0	4.5	1015.9
	37	8	51-24	174-07	8.20	7.78	32.85	32.81	9.0	8.0	0	4.5	1016.6
	38	8	50-28	174-44	7.80	7.65	32.90	32.90	6.4	6.2	100	4.5	1019.9
	39	9	49-31	172-58	6.60	6.33	33.01	32.99	7.0	6.8	100	4.5	1018.6
	40	9	48-48	171-08	6.00	5.89	33.10	33.10	7.6	7.0	unknown	2.0	1015.9
	41	10	48-39	169-11	5.80	5.42	33.00	33.00	6.4	6.2	100	2.0	1014.6
	42	10	48-56	167-18	6.10	5.82	33.12	33.12	7.0	6.6	100	4.5	1011.5
	43	11	49-09	165-46	6.20	5.95	33.13	33.13	7.2	6.8	100	4.5	1010.6
	44	11 12	49-35	164-05	6.80	6.57	33.15	33.17	6.8	6.6	100	4.5	1014.6
	Mean				6.77	6.53	33.06	33.05	7.4	6.9	75	3.0	1016.7

$\bar{\theta}_{w\cdot 0-10m}=6.65$°C $\quad \theta_{w\cdot 0}-\theta_{w\cdot 10}=0.24$°C $\quad \bar{\theta}_\alpha-\bar{\theta}'_\alpha=0.5$°C $\quad e_\alpha=9.38$ mb

$\bar{S}_{0-10}=33.05$‰ $\quad \sigma_t=25.94$ $\quad e_w=0.98\,(e_{\theta w\cdot \max})=9.59$ mb

* θ'_α...Temperature observed by wet valve.

Observations of Ice Crystals Formed on Sea Surface*

By

Yosio SUZUKI**

Introduction

It is well known that at the beginning of the freezing process of sea water small ice crystals are formed on its surface. Observations of such ice crystals in laboratory have been made by several authors using microscopic photography (Fukutomi, Nagashima and Kusunoki 1949) and shadow photography (Arakawa and Higuchi 1954; Arakawa 1955). However, as far as the present writer is aware, photographs of ice crystals on natural sea surface have not been published up to now.

In the course of the Sea Ice Expedition at Monbetsu (on the Okhotsk Sea coast of Hokkaido) in February 1955, the present writer tried to take photographs of ice crystals *in situ* conditions using polarized light source and polaroid filter. In this short report, the method and results of that attempt will be described.

Method

During the stay of the Expedition at Monbetsu, the inner harbor was covered with fast ice about 30 cm thick; the air temperature was always under 0°C. When a hole in this fast ice was made and the water surface was cleared of ice particles with a net, there soon began to appear newly formed ice crystals, which were so faintly visible in the day light, that special devices were needed for detailed observations.

The apparatus used is shown schematically in Fig. 1. Leica type camera (Canon Model IVS-b) was used with a reflex-box.

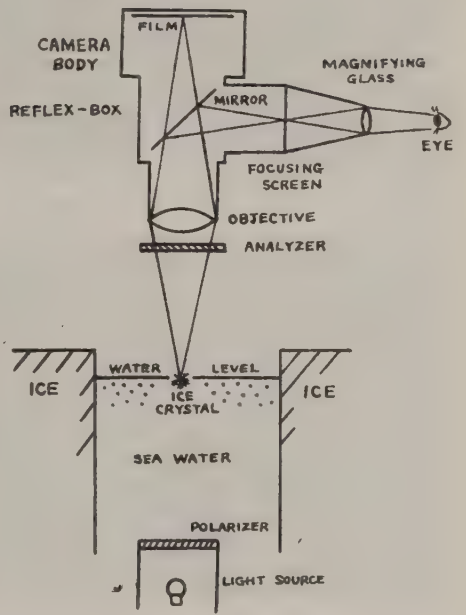


Fig. 1

Cameras of any type may be used but as the surface of sea water is not quite calm, focusing may be very difficult for cameras other than reflex type. Using a 50 mm objective and reflex-box, the enlarging ratio is about 1.4. The distance between the objective and the film must not be changed so as to keep the enlarging ratio constant. Focusing is done by moving up and down the camera body and the objective together. A water-proof flash light using 3 v dry cell is submerged about 10 cm under water as the light source. Using an $f=1.8$ objective, an exposure of 1/25 second and 1/2 second are adequate for panchromatic film (sensitivity: ASA 100) and color film (ASA 12), respectively. The use of color film is recommended because of its stronger stereographic effect.

* Contribution No. 281 from the Institute of Low Temperature Science. Received Aug. 3, 1955.

** The Institute of Low Temperature Science, Hokkaido University.

Results

No systematic observations were made in the winter of 1955. Only some 20 series of observations were made on the 22nd Feb. from 17-00 to 18-30 and on the 23rd from 17-30 to 19-00; about 60 photographs were taken, including some in color.

Figures 2-5 and 6-9 are two series showing the process of the crystal growth. The time intervals between the first and the last photographs of each series are about 10 minutes. Both series contain dedritic crystals but they are rather rare. Amorphous, needle and plate crystals are more frequently observed as shown in figures 10 and 11.

Some authors have reported about disc crystals. The writer observed round images out of focus moving rather rapidly. They might be disc crystals in the water.

Figures 12 and 13 are photographs of so-called ice rinds. Black circles in figure 12 are air bubbles.

Conclusion and Acknowledgment

Plans for systematic observations are now under consideration and preparation. It is expected that they will be carried next winter and also in the laboratory. Reports will be made later.

Finally, this work is partly supported by the Scientific Research Fund of the Ministry of Education, to the directors of which gratitude is expressed.

References

(1) Arakawa, K. and K. Higuhci 1954: On freezing process of aqueous solutions. Teion-Kagaku (Low Temperature Science), *Ser. A*, **12**, 73-86. (In Japanese with English résumé p. 80)

(2) Arakawa, K. 1955: The growth of ice crystals in water. J. of Glaciology, **2**, 17, 463-464.

(3) Fukutomi, T., T. Nagashima and K. Kusunoki 1949: Study of sea ice (2nd report); On the formation of ice crystals in sea water and the texture of sea ice. *ibid.*, **2**, 73-76. (In Japanese)

Appendix

Data of photographs and temperature of the air and the water.

Figures 2-5, 12, 13 Taken on 22nd February, 1955.
Figures 6-11 Taken on 23rd February, 1955.

Date and Time		Air Temperature (3 m above the ice)	Water Temperature (50 cm under the ice surface)
22nd	17.00	−1.5°C	−1.8°C
	18.00	−1.8	−1.8
	19.00	−2.3	−1.7
23rd	17.00	−1.8	−1.8
	18.00	−2.2	−1.8

(Chlorinity of the water was about 13‰)

Fig. 2

Fig. 3

Fig. 4

Fig. 5

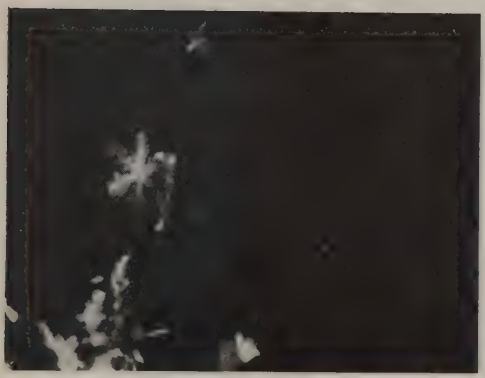
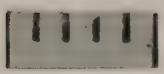


Fig. 6

Fig. 7

5 mm



Scale

Fig. 8

Fig. 9

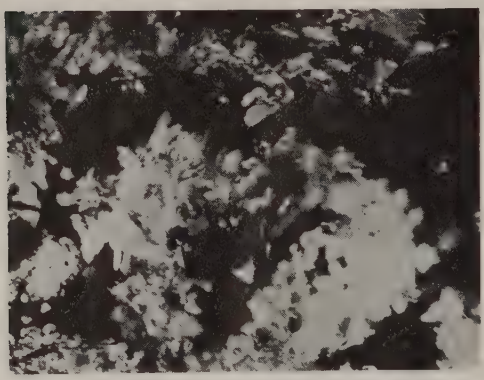


Fig. 10

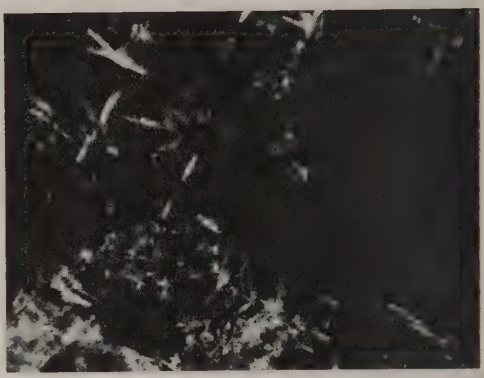
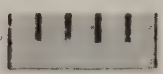


Fig. 11

Fig. 12

Fig. 13

5 mm



Scale

溶在酸素の年変化に関する一考察*
第二報

永井正男**

A Note on the Annual Variation of Dissolved Oxygen in Sea Water. Part 2.

By
Masao NAGAI

Abstract: In the present paper, the author has estimated the values of ΔQ_E, considering that the values of super-saturation in dissolved oxygen *in situ* were almost identical to the net production of oxygen by marine organisms. These results have been shown in Table 7 and Figure 5. On these results, the author has concluded that the dominant factor for the seasonal variation of dissolved oxygen may not be the exchange of oxygen across the sea surface (ΔQ_E), but the changes of solubility caused by the changes of water temperature. The seasonal variation of dissolved oxygen shows negative correlation, in which the coefficient of correlation is −1.0, with the seasonal variation of water temperature in Gotonada.

Thus, the net oxygen production resulting from biological activity is estimated about one-half of the magnitube of variation in dissolved oxygen caused by the changes of solubility.

§1. はしがき

著者は第一報において，(4) 式に示された Redfield の Exchange coefficient E は，Dorsey によつてヘンリーの法則から導き出された (5) 式の Entrance coefficient α に等しいことを証明し，(4) 式は海面における酸素の溶解機構については適用出来るが，海面内部の酸素量の変化については全く適用出来ないことを指摘した。第二報においては溶在酸素の年変化に関する著者の考察を紹介する。

§2. 現場飽和度と ΔQ_E

海洋学において通常用いられる気体の飽和度は，その時の水温・塩分における海水の気体飽和量と実際に存在している気体の 0°C・1 気圧における溶解量との比を百分率であらわしたものである。

ところで一般に気体の溶解度は次式で与えられる。即ち

$$S=\frac{v_g}{v_l} \tag{11}$$

こゝに S はオストワルド溶解度とよばれ，v_g は v_l なる容積の液体中に溶解した気体のその時の圧力及び温度における容積である。故に V_l 中に溶解した気体の質量を m として，この気体には理想気体の法則が適用出来るものとすれば，

$$m=\kappa' p v_g \tag{12}$$

こゝに p はその時の気体の圧力，k' は $m\cdot p\cdot v_g$ 等の単位に関係する恒数である。ところでヘンリーの法則によれば単位容積の液体に溶解する気体の質量は圧力に比例するから

$$\frac{m}{v_l}=\kappa p \tag{13}$$

こゝに k は比例恒数である。故に (12) と (13) の両式から

$$\frac{v_g}{v_l}=\frac{m/\kappa' p}{m/\kappa p}=\frac{\kappa'}{\kappa}=\text{一定} \tag{14}$$

即ちヘンリーの法則が成立するならば，溶解度 S は圧力の如何に関わらず一定温度に於ては一定でなくてはならない。

次に v_g を v_l なる容積の液体中に溶解した気体の 0°C·1 気圧のときの容積とする，とブンゼン吸収係数 B は次式で与えられる。

* 1955 年 7 月 14 日受理
** 長崎海洋気象台，Nagasaki Marine Observatory

$$B=\frac{v_g}{v_t} \qquad (15)$$

故に同じ気体の t°C·1 気圧のときの容積を v_g' とすると

$$v_g'=v_g(1+\alpha t) \qquad (16)$$

こゝに α は気体の膨脹係数 $\frac{1}{273}$ である。よつて (16) を (15) に代入すると

$$B=\frac{v_g'}{(1+\alpha t)v_t} \qquad (17)$$

然るに (14) 式により圧力の如何に関わらず, v_g/v_t は一定であるから, v_g' の 代りに v_g とすることが出来る。即ち

$$B=\frac{v_g}{(1+\alpha t)v_t}=\frac{S}{1+\alpha t} \qquad (18)$$

即ち気体の法則が成立するならば, 0°C·1 気圧における吸収係数 B に $1+\alpha t$ を乗ずるときオストワルドの溶解度を与える。

よつていま現場飽和度をオストワルドの溶解度 S とその時の気体の飽和量 S' との比の百分率としてあらわすものと定義するならば, それは (18) 式より通常我々が海洋学で用いる所の 0°C·1 気圧におけるブンゼン吸収係数 B に $1+\alpha t$ を乗じたものとして計算される。

Table 6. Seasonal variations of the oxygen content and saturation values *in situ.* in the surface layer of Gotonada

Depth	Factor	Month 1	2	3	4	5	6	7	8	9	10	11	12
0m	0_2	5.49	5.48	5.51	5.59	5.47	5.22	4.93	4.99	4.73	4.82	4.92	5.33
	S	5.84	5.79	5.84	5.95	5.85	5.63	5.35	5.49	5.17	5.25	5.32	5.68
	$0'_2$	5.65	5.81	5.71	5.62	5.46	5.23	5.12	4.77	4.95	5.03	5.19	5.59
	$S-0_2'$	0.19	−0.02	0.13	0.33	0.39	0.40	0.23	0.72	0.22	0.22	0.13	0.09
	S_s	103	100	102	106	107	108	104	115	104	104	103	102
5m	0_2	5.46	5.42	5.54	5.68	5.51	5.25	4.86	5.06	4.71	4.81	4.87	5.29
	S	5.80	5.73	5.87	6.04	5.90	5.66	5.27	5.57	5.14	5.24	5.27	5.64
	$0'_2$	5.65	5.78	5.75	5.64	5.47	5.26	5.12	4.80	4.95	5.04	5.19	5.58
	$S-0'_2$	0.15	−0.05	0.12	0.40	0.43	0.40	0.15	0.77	0.19	0.20	0.08	0.06
	S_s	103	99	102	107	108	108	103	116	104	104	102	101
10m	0_2	5.44	5.43	5.53	5.58	5.47	5.26	4.90	5.03	4.72	4.85	4.85	5.28
	S	5.78	5.74	5.86	5.93	5.85	5.67	5.31	5.53	5.15	5.28	5.24	5.63
	$0_2'$	5.65	5.78	5.75	5.64	5.45	5.27	5.15	4.83	4.96	5.04	5.19	5.57
	$S-0_2'$	0.13	−0.04	0.11	0.29	0.40	0.40	0.16	0.70	0.19	0.24	0.05	0.06
	S_s	102	99	102	105	107	108	103	114	104	105	101	101
25m	0_2	5.40	5.42	5.57	5.78	5.53	5.36	5.09	5.11	4.78	4.88	4.90	5.28
	S	5.74	5.73	5.90	6.14	5.91	5.77	5.50	5.61	5.22	5.31	5.30	5.62
	$0_2'$	5.65	5.78	5.75	5.68	5.51	5.32	5.23	4.85	4.97	5.03	5.18	5.59
	$S-0_2'$	0.09	−0.05	0.15	0.46	0.40	0.45	0.27	0.76	0.25	0.28	0.12	0.03
	S_s	102	99	103	108	107	108	105	116	105	106	102	101
50m	0_2	5.43	5.44	5.56	5.58	5.37	5.48	5.21	4.92	4.57	4.64	4.73	5.16
	S	5.77	5.75	5.89	5.91	5.71	5.85	5.59	5.37	4.98	5.04	5.10	5.49
	$0_2'$	5.68	5.79	5.75	5.70	5.64	5.49	5.38	4.98	4.99	5.08	5.19	5.60
	$S-0_2'$	0.09	−0.04	0.14	0.21	0.07	0.36	0.21	0.39	−0.01	−0.04	−0.09	−0.11
	S_s	102	99	102	104	101	107	104	108	100	99	98	98

Column 1. 0_2; the oxygen content in 0°C, 1 atm., c.c./L.

2. S; the oxygen content *in situ.*, $=0_2\times\left(1+\frac{t}{273}\right)$, c.c./L.

3. $0_2'$; the Fox's saturation values, c.c./L.

4. S-$0_2'$; the quantity of super-saturated oxygen *in situ.* the estimates of the net production in oxygen resulting from biological activity at the observed layer.

5. S_s; the percentages of saturation *in situ.* to the Fox's values, *i.e.*$=S/0_2'\times 100\%$.

Table 6 にはかくして求めた O_2 のオストワルド溶解度 S とその時の温度・塩分における Fox の飽和量 O_2' 及び現場飽和度 S_s の値が示してある。即ち現場飽和度は各層とも一年を通じてほとんど 100% 以上であり，ヘンリーの法則から導かれる Dorsey の

$$c/c_s = 1 - e^{-\beta \cdot 17/\alpha}$$

なる関係から推論される，水面の薄いフイルムにおける酸素の吸収は常に水面を飽和の状態に進行させるという結果が現実の問題として裏ずけられたことになる。ことに海面においては水面が静止している場合とは異なり，風浪による影響が考えられ，Table 5 に示した値よりは尚飽和に達するまでの時間は小さいであろうということがうなずかれる。

従つて Redfield が述べているように (4) 式の $(P-p)$ の値が，10 月～3 月 では正となり大気中から溶解されるが，4 月～9 月の 6 ケ月では負となつて同じ量だけ大気中え逃げて行くということは全く考えられないことになる。

ところでヘンリーの法則だけからでは，海面が常に 100% に保たれるであろうことはうなずけるが，どうして 100% 以上になるかという解答は与えられない。

即ちその時の水温と塩分における現場飽和量以上に酸素が溶解されているならば，それは生物活動による酸素の正味の生産量と考えてもよいであろう。

而して現場飽和度が 100% 以上の過飽和の酸素量は潜在不安定な状態にあるため，環境の変化即ち水温と塩分の変化による酸素の溶解度が変化する時，此等の潜在不安定な酸素量は大気中に逸散するという訳である。即ち Redfield の ΔQ_E は常にマイナスの値として表現されるべき量であり，$P-p$ の値によつて惹起されるべき量ではなく，水温と塩分の変化に起因する酸素の溶解度の変化をあらわす値である。

而して此のことは五島灘における Fig. 2 及び Fig. 3 を比較してみれば一層よくわかることであり，ΔQ_E と $P-p$ の値とはお互に無関係な値であつて，dQ/dt が $P-p$ に比例するという (4) 式は適用出来ないことがうなずかれる。即ち Redfield は (4) 式によつて Exchange coefficient なる常数を計算し，その値が負では小さく 2.8×10^6 c.c./m². atm. month であり，冬では大きく 13×10^6 c.c./m². atm. month で，それはおそらく冬においては季節風が強いため夏におけるよりも海面近くの状態が不安定で白波の現象を起し易く，大気中から海水中え溶けこむことを増大させるような役割を果しているといつているが，此れは全くの誤謬であつて，白波の現象は現場飽和度が 100% 以上の不安定な酸素量によつて惹起されるものであり，気泡が大気中え逸散する現場であると解釈される。

又 Redfield の求めた E の値が，Dorsey の α の値 0.017 c.c./min. cm.² atm. ($=7.3\times10^6$ c.c./month. m.² atm.) とオーダーの点で割合よく一致しているといつていることは，以上のことからもわかるように全く偶然の一致というべきで，Table 4 に示すように Redfield の求めた方法と全く同じ方法で求めた $E_1 \cdot E_2$ の値からすればわかるように (C_0 の値として Fox の飽和量に水蒸気張力の補正をほどこしたものを使つた場合の Exchange coefficient は E_1 で，又補正をほどこさない場合のそれを E_2 であらわす)，一番小さな値でも 3～4 月の 1×10^{-4} で 10^{-6} のオーダーにはなり得ない。もつともこれは五島灘の表層水における 50m 層までをとつた値で Redfield の場合におけるように 200m 層までの値をとつていないということに起因するものであるが，Redfield の Maine 湾における場合は Table 2 に示した Q_0 の値が 1×10^6 のオーダーであり，五島灘の場合は 0.25×10^6 のオーダーで明らかに前者の 1/4 となつている。即ち Redfield の方法で求めた E の値が Dorsey の α の値にオーダーの点で一致したということは，偶然考慮に入れた深さが Dorsey の α の値とオーダーの点で一致するような深さであつたということにしか他ならない。

又 Table 4 の $E_1 \cdot E_2$ の値から Redfield のいうように夏では小さく，冬では大きいという傾向が全く見られないことも当然の帰結として附記しておかねばならないだろう。

§3. 結果の考察及び吟味

では一体 Redfield の ΔQ_x, 即ち生物活動による酸素の正味の生産量はどのような形であらわされるのであろうか。

著者はさきに現場飽和度が 100% 以上の酸素量は，生物活動による正味の生産量と考えても差支えないであろうということを述べた。

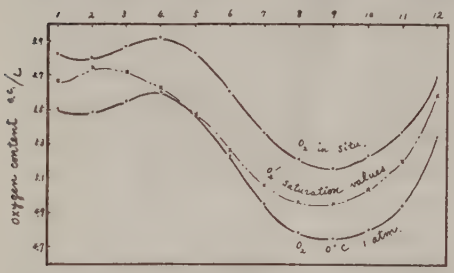


Fig. 4. Seasonal variations of the oxygen content under the condition of 0°C, 1 atm. and *in situ*., and $0_2'$, Fox's saturation values, at the surface layer of Gotonada.
(The curve was smoothed by hand.)

Fig. 4 には Table 6 に示した 0m 層の $0_2 \cdot S \cdot 0_2'$ の値を平滑した曲線を示してある。即ち $S-0_2'$ が 0m 層における生物活動による酸素の正味の生産量をあらわすと考える訳である。

所で今 Table 1 に示した 0_2 の値をみると，年変化の曲線は各深さともほとんど表面と同様であることがうかゞわれる。而して特に相違をみつけるならば，25m 層の 4・5・6 月の値がやゝ大きく，生物活動による酸素の正味の生産量の増加を示していることゝ，50m 層の 8 月以降においてやゝ小さな値を示していることである。又 $0_2'$ の年変化も 50m 層の 5・6・7・8 月でやゝ大きな値を示している他はほとんど一様で水温・塩分共閉鎖系の条件に適合した同一水系における年変化であることをよく表わしている。

Table 7. Estimations of $\Sigma(S-0_2')$ the net production in oxygen resulting from biological activity at the surface layer of the Gotonada.

Depth \ Month	1	2	3	4	5	6	7	8	9	10	11	12
0	0.19	−0.02	0.13	0.33	0.39	0.40	0.23	0.72	0.22	0.22	0.13	0.09
5	0.15	−0.05	0.12	0.40	0.43	0.40	0.15	0.77	0.19	0.20	0.08	0.06
10	0.13	−0.04	0.11	0.29	0.40	0.40	0.16	0.70	0.19	0.24	0.05	0.06
25	0.09	−0.05	0.15	0.46	0.40	0.45	0.27	0.76	0.25	0.28	0.12	0.03
50	0.09	−0.04	0.14	0.21	0.07	0.36	0.21	0.39	−0.01	−0.04	−0.09	−0.11
$\Sigma(S-0_2')$	0.65	−0.20	0.65	1.69	1.69	2.01	1.02	3.34	0.84	0.90	0.27	0.13

Column. 1. S; the oxygen content *in situ*., $=0_2 \times \left(1+\frac{t}{273}\right)$ c.c./L.

2. $0_2'$; the Fox's saturation values, c.c./L.
$=10.291-0.2809\,t+0.006009\,t^2-0.0000632\,t^3-\mathrm{Cl}(0.1161-0.003922\,t+0.0000631\,t^2)$
here, t is the water temprature in °C and Cl is the chlorinity in ‰ *in situ*., .respectively.

3. $S-0_2'$; the quantity of super-saturated oxygen *in situ*, at the observed layer shown in Table 6.

4. $\Sigma(S-0_2')$; the estimated values of the net production in oxygen resulting from biological activity, which represent the integrated values of $S-0_2'$ in the layers 0～50 m. Unit: $\times 10^4$ c.c./m².

よって今 t °C・1 気圧における現場溶解量 S と溶解度の変化に起因する $0_2'$ の差即ち $S-0_2'$ を計算してみることにしよう。Table 7 には此等の値が示してあるが，一見していゝ得ることは非常によくそれらの年変化が一致しているということである。即ちそれらの値の平均値から 50m 層に至るまでの $\Sigma(S-0_2')$ を求めれば近似的には生物活動による酸素の正味の生産量をあらわすと考えても差支えない。

しかるとき Table 7 に示した 50m 層における 9・10・11・12 月の値がマイナスの符号を示しているのは，此等の月では最早生物活動による酸

素の生産量よりも消費量の方が大きいことを物語っていると考えればよい。

しかしながら $S-O_2'$ の値は何等の吟味もなしにそのまゝすぐ生物活動による酸素の正味の生産量をあらわすとはいゝ得ない。即ち O_2' の値は水温と塩分とにもとずく量であるから，$S-O_2'$ の値も水温と塩分の変化に無関係だとはいえないということである。而してこの最も適切な例が五島灘では 8 月においてあらわれている訳である。(Table 1 を一見してもわかるように 8 月における O_2' は 7 月・9 月のそれに比べ特別に小さな値を示している。これは 8 月における水温が 7 月・9 月の値に比べ急激に上昇していることが原因である。即ち 8 月における $S-O_2'$ は水温の影響が大きいため生物活動による酸素の正味の生産量をあらわすとはいゝ得ない。)

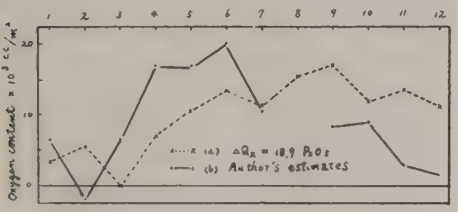


Fig. 5. Estimations of the net production of oxygen resulting from biological activity at the surface layer of Gotonada.
(Dotted line represents the values calculated from the formula $\Delta Q_z = 18.9\ P_2O_5$, based on the Redfield's estimation.)

では一体 $S-O_2'$ はどの程度まで生物活動による酸素の正味の生産量をあらわすものであろうか。Fig. 5 には Redfield による (2) 式からの ΔQ_z の推定曲線 (a) と，Table 7 の $\sum S-O_2'$ による著者の推定曲線 (b) とが示してある。即ち前者は Fig. 2 にも示してあるように P_2O_5 の変化量より求めた曲線で，3 月に Min があらわれ，9 月に Max があらわれている。又後者においては Min は 2 月にあらわれており，Max は 5 月の下旬と想像される。($\sum S-O_2'$ よりすれば 2 月の値はマイナスであるが，2 月は 1 月・3 月等から推定される値よりも O_2' が大きいためで，生物活動による酸素の正味の生産量は (b) 曲線の変化から考えてもわかるように必ずしもマイナスと考えない方が妥当である。又 $\sum S-O_2'$ からすれば 5 月より 6 月の方が値は大きいが，これは水温の上昇による影響で S が大きくなることゝ O_2' が小さくなることのために其の差は 5 月よりも大きくあらわれていると考えた方が妥当と思われる。)

而して此等 (a)，(b) の推定曲線を比較していゝ得ることは年変化の形はまるで異なるが，此等の値は両者共オーダーの点では一致しており，(b) の Max は (2) 式の係数を大略 22 としたときの値である。即ち (b) の方がやゝ大きな値となつてあらわれているが，これは (b) の曲線が生物活動による正味の生産量と水温の上昇による影響とがプラスされていることを考えればよく一致しているということがいえるであろう。

所で著者の考えに従えば海水中における溶在酸素の年変化は，閉鎖系の条件が満足されるならば，t°C・1 気圧における現場溶解量 S を考えることによつて，$S-O_2'$ の値が生物活動による酸素の正味の生産量をあらわし，O_2' は水温・塩分の変化による溶解度の変化をあらわして，渦動拡散による Max の溶解量として解釈された。

即ち海水中における溶在酸素の年変化は，もしも生物活動による酸素の正味の生産がないとすれば，水温・塩分によつて定まる O_2' の値以上には溶存することはなく，もしも渦動拡散が完全におこなわれていると仮定すれば，現場飽和量は一様に 100% になることが考えられる。即ち O_2' は水温と塩分を測定すれば，Fox の式から求まる量であり，生物活動による酸素の正味の生産量は現場飽和量 S から O_2' を引いた値であるから，Fig. 4 に示した各曲線からもわかるように

$$\sum(S-O_2')+\sum O_2'=Q_z+Q_E=Q_0 \qquad (19)$$

なる関係が成立する。

而してこの式は Redfield の出発点をなした (1) 式と全く同じ形のものである。しかしこれらの記号のあらわす内容は既に述べて来たように全く違つたものであることは注意を要する。即ち Q_z は生物活動による酸素の正味の生産量であるが，Redfield では P_2O_5 の消費量から計算される値であり，著者の場合は現場飽和度が 100% 以上

の値によつて推定される量である。又 Q_E は生物活動には無関係な物理的要因による量で，Redfield では $(P-p)$ の大気中と海水中の酸素分圧の差によつて海面を出入する酸素量であつたが，著者の場合は水温と塩分の変化による溶解度の変化をあらわす，Fox の飽和量そのものゝ値であり，Q_0 は何れも海水中の溶在酸素量で Q_x+Q_E によつて定まる量であるが，Redfield では 0°C・1 気圧における通常の値であるのに反し，著者の値は t°C・1 気圧における現場飽和量をあらわしている。

さらに Redfield においては Table 2 に示した $\Delta Q_0 \cdot \Delta Q_E \cdot \Delta Q_x$ の値について，どうして ΔQ_0 がマイナスの量として規定されねばならないのかその理由が判然としない。即ち ΔQ_0 は単に 4 月の Max の値との差としてあらわれているだけで，マイナスでなければならないという必然性は何処にもなく，又 ΔQ_0 がマイナスであることは現実の問題としてどういうことかゞ解釈出来ない。又 ΔQ_0 がマイナスでないとすれば，ΔQ_E もマイナスにはならないから (1) 式の関係より $|\Delta Q_0| > |\Delta Q_E|$ となり，Table 2 におけるような $\Delta Q_0 < |\Delta Q_E|$ なる関係は成立しない。又 ΔQ_E がプラスであるということは空気中より溶解されるということで現実的には誤謬であることがたゞちに指摘されるであろう。

§4. むすび

以上において溶在酸素の年変化に関する著者の考察を，Redfield の誤謬を指摘しながら述べて来たのであるが，従来は水温と塩分の相違にともなう酸素の溶解度の変化というものは考察されずに来たようである。即ち溶在酸素の変化量はそのまゝすぐに生物活動による酸素の正味の生産量と考えた訳である。しかしながら以上の議論からもわかるように溶解度の変化にともなう酸素の変化量は一番大きく，ほとんど年変化の曲線を決定すると考えてよい。即ち五島灘においては生物活動による酸素の正味の生産量の大略 2 倍にあたると考えられる。(Fig. 4 より生物活動による酸素の正味の生産量は 5 月及び 6 月の Max で大略 0.4 c.c./L と推定されるが，溶解度の変化による酸素の変化量は 2 月と 8 月のほゞ 10°C の水温差に対し 0.8 c.c./L となつている)。

終りに以上を要約すれば，

(1) 溶在酸素の年変化を支配する第一の要因は，Redfield が述べているように ΔQ_E 即ち海面を通過する酸素の量ではなく，水温と塩分の変化による溶解度の変化で，五島灘における溶在酸素の年変化は水温の年変化と −1.0 の逆相関が見られる

(2) 0m 層即ち海面における溶在酸素の現場飽和度は一年を通じて 100% 以上の値を示し，Redfield が述べているように (4) 式に示す $(P-p)$ の値が，10 月～3 月では正となり大気中から溶解されるが，4 月～9 月では負となつて同じ量だけ大気中え逃げて行くということは全く考えられないことになる。

(3) 又現場飽和度が 100% 以上の過飽和の酸素量は潜在不安定な状態にあるため，環境の変化即ち水温と塩分の変化による酸素の溶解度が変化する時，此等の潜在不安定な酸素は空気中に逸散する。

(4) 而してこの現場飽和度が 100% 以上の過飽和の酸素量は水温・塩分の急激な変化がない限り，大略生物活動による酸素の正味の生産量をあらわすと考えても差支えない。

(5) 然るとき生物活動による酸素の正味の生産量と溶解度の変化にもとずく酸素の変化量との比は大略 $1/2$ の値を示す。

文　献

(1) Redfield, A. C., The exchange of oxygen across the sea surface. J. Mar. Res., VII p. 347～361, 1948.

(2) Dorsey, N. E., Properties of ordinary water substance. Reinhold, New York. p. 552-555, 1940.

(3) 寒作新六：物理化学測定法（共立出版社）1st ed. p. 198-199.

(4) 三宅泰雄：海中における気泡発生の可能性とその限界（海水の泡沫性に関する研究，その 1）海と空 Vol. 24, No. 4, p. 121～134, 1944

(5) 三宅泰雄，阿部友三郎：海水泡沫の大さと安定度について（海水の泡沫性に関する研究，その 2）海と空 Vol. 24, No. 9, p. 291～295, 1944

(6) 松平康雄：海水中の酸素含量に対する実験 海と空 Vol. 12, No. 5, p. 91～101, 1932

Analysis of Water Masses by Distributions of the Microplankton (1)*

Distributions of the Microplankton and their Relation to Water Masses in the North Pacific Ocean in the Summer of 1954

By

Ryuzo MARUMO**

§ 1. Preface

There exists an eastward current system in the sea area around 40°N in the North Pacific Ocean. This current is considered to be formed in its western part by both the Kuroshio warm water and the cold water of northern higher latitudes, which is called the Oyashio in the northeast area of Japan, and finally to be connected with the California Current which streams southward along the coast of North America (Fig. 1). In the present paper, characters

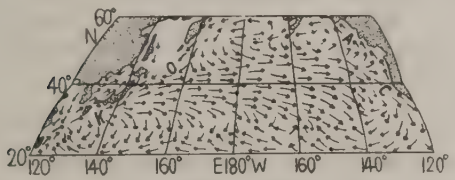


Fig. 1. Surface currents of the North Pacific Ocean (after Sverdrup and others). K: Kuroshio Current, O: Oyashio Current, A: Alaska Current, C: California Current.

of this current system, especially its origin, are obviously connected to distributions of surface microplankton, and additionally some results are obtained also about the microplankton in the areas of 20 to 30°N.

Plankton samples were collected on board the *Nihon-maru* of the Navigation Training Establishment at stations on the N-(May 2-June 2) and S-lines (June 13-August 2) on her training voyage between Tokyo and Los Angeles (Fig. 2). Surface water (330cc) was drawn up with a bucket. This water was reduced to 15cc by decanting with a siphon and the 15cc water was again

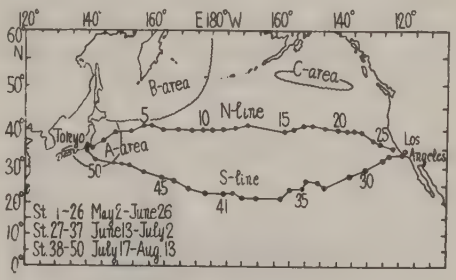


Fig. 2. Observational stations.

reduced to several drops with a centrifuge. Plankton organisms in these drops were identified. The method used in this study is already detailed by Marumo (1954). The cell number means that of diatoms in 1 liter of surface sea water in this paper.

§ 2. Biological characters of surface water on the N-line

1. *An outline of distributions of the microplankton*

At station 1 water temperature was 18.1°C, the highest on this line (Fig. 3). The diatom population of this station was fairly large, $4 \cdot 10^3$ cells to the liter, being composed of *Leptocylindrus danicus*, *Rhizo-*

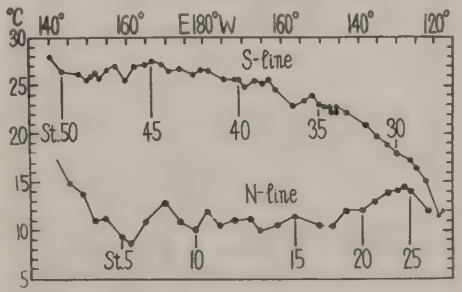


Fig. 3. Distributions of surface water temperature.

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** Oceanographical Section, Central Meteorological Observatory

solenia stolterfothii, Subgenus *Hyalochaete* (*Chaetoceros affinis*, *Ch. didymus*, *Ch. compressus*, *Ch. lorenzianus*), *Eucampia zoodiacus*, etc., which indicate a neritic Kuroshio water (Figs. 4 and 5).

Between stations 2 and 8 the population was small, less than 10^2, and contained only a few species such as *Coscinodiscus excentricus*, *Cos. stellaris* and *Chaetoceros concavicornis*. The community like this is often distributed in the oceanic Oyashio water in the northeast area of Japan and is found in the water of about 5°C by

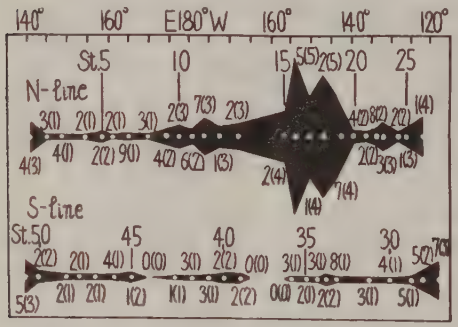

Fig. 4. Distributions of number of diatom cells in 1 liter of surface sea water. The figures attached in parentheses mean the power of 10, for instance, 4(3) denotes 4000.

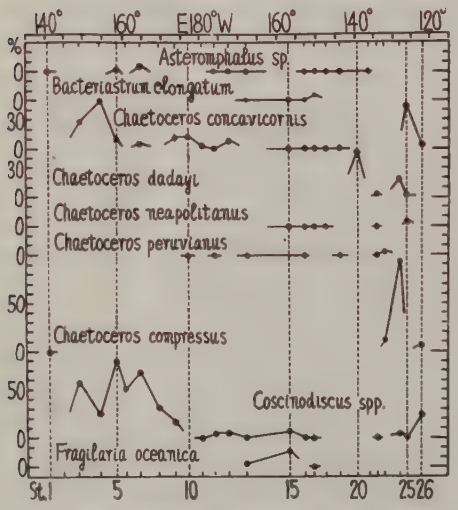

Fig. 5-a. Percentages of the cell number of important species to the total diatom cells.

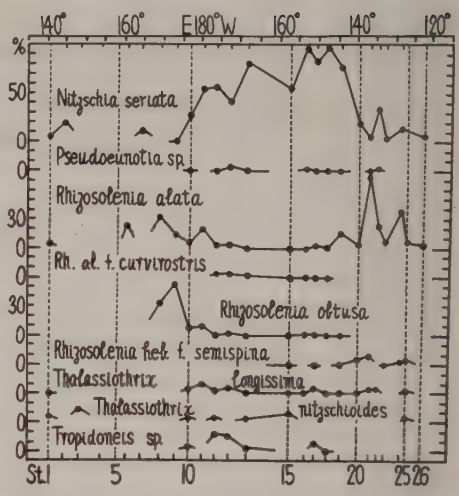

Fig. 5-b. Percentages of the cell number of important species to the total diatom cells.

Marumo and others (1955).

The area between stations 8 and 10 is also characterized by cold water, because *Rhizosolenia alata*, *Rh. obtusa* and *Nitzschia seriata* were dominant in addition to the species found between stations 2 and 8, and because cold-water tintinnids such as *Parafavella* spp. and *Acanthostomella norvegica* appeared frequently.

From st. 11 toward st. 19 the cell number increased gradually and the densest region of 10^4 to 10^6 was found in the area between stations 15 and 19. The number reached the maximum, $5 \cdot 10^5$, at st. 16. The percentage of *Nitzschia seriata* was highest, 82%, between stations 15 and 19, on an average. And in this area the number of diatom species was also largest, being newly accompanied by *Fragilaria oceanica*, *Pseudoeunotia* sp., *Rhizosolenia alata* f. *curvirostris*, *Rh. hebetata* f. *semispina*, *Thalassiothrix longissima*, *Th. nitzschioides*, *Tropidoneis* sp., *Bacteriastrum elongatum*, *Chaetoceros atlanticus* v. *neapolitana* and *Ch. peruvianus*.

Between stations 20 and 25 the population decreased suddenly to 10^2–10^3 and the community shows a warm-water character. *Nitzschia seriata* was still the leading species, while the distinctly cold-water

species such as *Fragilaria oceanica*, *Rhizosolenia alata* f. *curvirostris*, *Rh. obtusa* and *Tropidoneis* sp. dissapeared perfectly. It may be an interesting fact that a typical warm-water species, *Chaetoceros dadayi*, was detected in this area, especially in a high percentage of 42% at st. 20. This species is found in the Kuroshio of high temperature mainly in summer and autumn in the sea around Japan. Moreover, warm-water species such as *Trichodesmium* spp., *Ceratium inflexum*, *Cer. massiliense*, *Cer. macroceros* and *Amphorella quadrineata* were also found, and on the other hand a neritic species, *Chaetoceros compressus*, reached 93% at st. 23. It may, therefore, be concluded that this area was affected by some neritic warm water.

At st. 26 the population increased again to $3 \cdot 10^3$, forming a rather cold and neritic community which contain *Eucampia zoodiacus*, *Skeletonema costatum*, *Chaetoceros compressus* and *Ch. radicans*.

2. *The origin of the eastward current water inferred from distributions of the microplankton*

It may be easily drawn from the above description that the area between stations 2 and 19 is characterized by cold water from the view point of microplankton communities. And further discussion shall be done on the origin of this current water, making reference to distributions of the microplankton in the Kuroshio and Bering waters.

Surface plankton samples were collected in the A-area of Fig. 2 on board the ships of the Japanese marine meteorological observatories (the Central Meteorological Observatory, the Hakodate and Kobe Marine Observatories) in the same way as in the present survey in the period May to June 1954. In this season the characteristic species was subgenus *Hyalochaete* (*Chaetoceros lorenzianus*, *Ch. affinis*, *Ch. didymus*, *Ch. messanensis*, *Ch. curvisetus*, *Ch. compressus*), *Rhizosolenia stolterfothii*, *Leptocylindrus danicus*, etc., in the Kuroshio water of 20 to 24°C. As these species were contained also in the waters with temperatures of 10 to 20°C, which are considered to be due to the mixing of the Kuroshio and Oyashio, it may be most probable that they would be distributed also in a range of 10 to 13°C in the area between stations 2 and 19. However, none of the Kuroshio species stated above was found in this area at all and only cosmopolitan species, *Thalassiothrix nitzschioides* and *Nitzschia seriata*, were present in common in both the areas. The fact that characteristic species indicating the Kuroshio water were not found between stations 2 and 19 shows that the Kuroshio water did not come distinctly into the eastward current system.

Fortunately we have some data useful to know a typical community of cold-water diatoms in this season. Surface plankton samples were collected in the B-area south of the Kamchatka Peninsula on board the *Tenyo-maru* of the Tokai Regional Fisheries Research Laboratory in the said way in the period June to August 1954. The water temperature was less than 10°C in this area. With the exception of a neritic community composed of *Chaetoceros debilis*, *Ch. compressus*, etc. in a part along the Kuril Islands, the characteristic species were *Chaetoceros concavicornis*, *Ch. convolutus* and *Nitzschia seriata* which were accompanied by *Chaetoceros atlanticus*, *Corethron hystrix*, *Denticula* sp., *Rhizosolenia alata*, *Rh. hebetata* f. *semispina*, *Rh. hebetata* f. *hiemalis*, *Rh. obtusa*, *Tropidoneis* sp. and *Parafavella* spp. as secondary components. Only one of these species, *Rhizosolenia hebetata* f. *hiemalis*, was not detected in the area between stations 2 and 19.

The result demonstrates that the water between stations 2 and 19 has originated from the oceanic cold water in the Bering Sea or its neighbourhood. Especially, such species as *Chaetoceros atlanticus*, *Rhizosolenia alata* f. *curvirostris*, *Rh. obtusa*, *Fragilaria oceanica*, *Pseudoeunotia* sp., *Tropidoneis* sp., *Paarfavella* spp. and *Acanthostomella norvegica* must be regarded as excellent indicators of cold waters. Some consideration is necessary for the appearance, though small in quantity, of *Bacteriastrum elongatum* at stations 14 to 17

and of *Chaetoceros atlanticus* v. *neapolitana* at stations 15 to 18. Because these species were found in the Kuroshio water but not found in the Bering water, being regarded to be characteristic of warm water, it may be right to say that the water between stations 14 and 18 has previously mixed with some warm water somewhere. However, these species were not at all found in the west area of st. 13 on the N-line.

Allen (1927) reports that the leading species were *Chaetoceros socialis*, *Ch. debilis*, *Skeletonema costatum* and *Eucampia zoodiacus* in the C-area of Fig. 2, where the water temperature was about 10°C in October 1923. If these species were present in the Alaskan water also in June 1954, they are expected to come into the water west of st. 15. However, these typical neritic species were not present here at all in the present survey.

Now, it may be concluded that the water between stations 2 and 19 originates mainly from the Bering Sea or from its neighbourhood and is hardly mixed with the Kuroshio, being free from the influence of neritic waters. Although considerable numbers of warm-water and cold-water species (the Kuroshio and Oyashio species in the sea around Japan) are possible to exist together in a range of 10 to 13°C, to say nothing of cosmopolitan species, only cold-water and cosmopolitan species were found in this case. This fact may suggest that distributions of the microplankton are not always dependent upon the water temperature but sometimes upon original characters of the water. It can not be concluded from the distributions of water temperature that this area of 10 to 13°C is characterized by cold water in origin. On the N-line, plankton collections were made at large intervals of 200 nautical miles, but the cell number and community of diatoms changed in succession geographically. This result seems to show that plankton diatoms are fairly uniformly distributed in high seas and that water masses can be effectively analysed by the same way as in this study.

The temperature in the area between stations 20 and 25 is 12 to 15°C and slightly higher than in its western area. The appearance, however, of many warm-water species in the former area is not reasonably explained by such a small difference in temperature in the two areas, but it seems to be due to some other biological characters of the water itself. We have not sufficient plankton data to assure the origin of the water mass, but the following conjecture may be possible. Namely, a northward warm current along the coast of North America, which is shown in Fig. 1, reached here through some course and the existence of *Chaetoceros compressus* show the mixing with neritic waters.

§ 3. Biological characters of surface water on the S-line

This area is situated between 20 and 30°N and diatoms were remarkably scanty, less than 10^2 cells to the liter. The community, however, was neritic and the cell increased to $2 \cdot 10^2$–$7 \cdot 10^3$ at coastal stations 27, 28, 50 and 51. Only one of 25 samples on the S-line did not involve any plankton organism. Collected diatoms were too small in number to obtain a perfect view of their community, but generally *Rhizosolenia styliformis* was most prevalent, being accompanied by *Hemiaulus hauckii* and *Rhizosolenia calcar avis* as the secondary components, and sometimes *Chaetoceros tetrastichon*, *Climacodium frauenfeldianum*, *Trichodesmium* spp., etc. were found. Allen and Lewis (1927) also report that plankton organisms were very scarce in surface waters of lower latitudes in the North Pacific Ocean.

Table 1. Average individual numbers of copepods and tintinnids in 1 liter of surface sea water.

	copepods	tintinnids
N-line	45	3.1
S-line	3.0	1.5

The average individual number of copepods (including all stages) and tintinnids which were distributed abundantly next to diatoms is shown in Table 1. Copepods are remarkably excellent on the N-line but

tintinnids do not show any notable distinction in number on the two lines.

§ 4. Summary

1. The number of diatom cells increases from 10^2 to 10^6 to the liter from st. 2 toward st. 19 on the N-line. The community of this area indicates that this water has originated from the Bering Sea or from its neighbourhood and is hardly affected by the Kuroshio and neritic waters.

2. The cell number is 10^2 to 10^3 between stations 20 and 25 on the N-line. And the warm-water species in this water seems to have been contained in the northward current along the coast of North America.

3. The cell number is very small, less than 10^2, between stations 29 and 49 on the S-line. *Rhizosolenia styliformis* is most prevalent.

4. The cell number is 10^2 to 10^5 at coastal stations 1, 26, 27, 28, 50 and 51 and their community is characteristic of neritic water.

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References

Allen, W. E. (1927): Surface Catches of Marine Diatoms and Dinoflagellates made by U. S. S. "Pioneer" in Alaskan Waters in 1923, *Bull. Scripps Inst. Oceanogr.*, Tech. Ser., **1** (4).

Allen, W. E. and R. Lewis (1927): Surface Catches of Marine Diatoms and Dinoflagellates from Pacific High Seas in 1925 and 1926. *ibid.*, **1** (12).

Marumo, R. (1954): Diatom Plankton in the South of Cape Shionomisaki in 1953, *Oceanogr. Mag.*, **6** (3).

Marumo, R., F. Uyeno, Y. Kawarada, F. Ogawa and T. Akiyama (1955): An Example of the Analysis of Water Masses by the Microplankton, *Journ. Meteorol. Res.* (in Japanese) (in printing)

録　　事

最近の海洋学関係図書

最近 2, 3年に国内で出版された図書を紹介する。

(1) 日本海洋学会編：「海洋観測指針」(昭和 30 年 3 月)，頒価 600 円，附録常用表共，神戸海洋気象台の「海洋観測法」を気象台海洋関係者が寄つて改訂したもので，海洋観測者の国内基準書として必携すべきものであろう。水産関係海洋調査にはこの外標識放流，魚介藻，卵，稚魚，ベントス等の標本採集，調査等が必要であり，特に海上作業には「海洋気象法」又は「船舶気象観測指針」も一しよにしたもう少し簡略なものがほしいと思われる。

(2) 小久保清治著：「浮游硅藻類」，日本学術振興会出版（丸善発売，定価 1000 円)，昭和 30 年 3 月。プランクトン学の権威，東北大学名誉教授小久保博士が半年の心血を注いだ大著である　各大学，研究所，官庁諸学校等に 1 冊備うべきこの方面の分類等に必要な基準書と思われる　動物プランクトン等についても待望される。

(3) 三宅泰雄著：「地球化学」，朝倉書店（定価 780 円) 昭和 29 年 7 月。内容は気圏，水圏，岩石圏に亘り，海洋化学についても最新の知識をまとめた好参考書である。気象研究所三宅博士の得意とする部門の研究を網羅された　水産生物に重点をおく Harvey 流の本もほしい。

(4) (a) 宇田道隆著：「海洋気象学」，天然社（定価 500 円) 昭和 29 年 6 月．表層の海洋学（海流，波浪，海氷等）と海上気象学（暴風，海霧等）を含み，航海水産の防災，特に漁業気象につとめ，初めてまとまつた海洋気象学書．

(b) なお天候変動についての快著，高橋浩一郎著：「動気候学」，岩波書店（定価 850 円)，昭和 30 年 3 月も推奨したい。石丸雄吉著：「雲の気象学」(北隆館，昭和 27 年) も好著である。

(c) 正野重方著：「気象力学序説」，岩波書店（定価 1000 円)，昭和 29 年 9 月。高級書で近代理論気象学入門。「海洋物理学」もこの程度の J. Proudman: Dynamical Oceanography (1953) に匹敵するものがほしい。

(5) 航海海洋学に関係して球面天文学の知識は必須のものである。中野猿人博士（気象研究所）はさきに「潮汐学」(古今書院) の大著を出したが，昭和 27 年 6 月「球面天文学」，(古今書院) の良書を出した。(定価 1200 円，684 頁)。

(6) 其他海洋学関係雑書名（普及書を含む)。

(a) 須田暁次編修：「うみ」，玉川こども百科（昭和 29 年 6 月)。

(b) 日高孝次訳（カーソン女史著)：「海，その科学とロマンス」，文芸春秋新社 (昭和 27 年 7 月)。

(c) 末広恭雄訳（クラーク女史著)：「銛を打つ女」，法政大学出版局 (昭和 29 年)。

(d) 宇田道隆著：「海」(増補版) 岩波新書，岩波書店 (昭和 28 年)。

(e) 松江吉行著：「海の科学」，金子書房 (昭和 28 年)。

(f) 三宅泰雄著：「海のなぞ」，牧書店 (昭和 27 年)。

(g) 宇野木早苗著：「海の波」，恒星社 (昭和 29 年)。

(h) 宇田道隆著：「海と魚—少国民のために」，岩波書店 (昭和 27 年訂補)。　その他。

こうして見ると「海洋地質学」の Ph. H. Kuenen: Marine Geology (1950) のような本がほしい。生物海洋学では Sven Ekman: Zoo-Geography of the Sea (1953) のような本で水産，漁場の色彩の強いものも未だない．海洋図集（海洋気象図集を含めて）も要求されている

(宇田道隆記)

Roger Revelle 博士及び Gordon Lill 氏の日本訪問　アメリカ合衆国スクリップス海洋研究所長 Dr. Roger Revelle 及びアメリカ合衆国海軍研究所地球物理部長 Mr. Gordon Lill の両氏は，本年実施される日・米・加協同北太平洋調査 (Nor Pac Expedition) 及び 1957～8 年実施予定の国際地球物理観測年 (International Geophysical Year) に関する準備打合のため，去る 7 月上旬日本を訪れ，約 10 日間滞在し，関係方面と交歓打合を行つた後無事帰国した。この間，7 月 2 日には東京大学理学部第 2 号館で両氏の学術講演会が催され，Revelle 博士は "北太平洋の三つの謎" と題し，主として海底地形学や海底地質学の立場から見た太平洋の海洋学について興味深い講演をなし，また Lill 氏は "国際地球物理観測年について" と題し，その沿革史と現状について詳しく説明した。また，7 月 6 日には，東海区水産研究所で両氏を中心とする "海洋学の水産業への応用に関するシンポジウム" が開かれ，Revelle 博士は特に合衆国太平洋側のイワシ漁業，マグ

ロ漁業を中心とする漁業上の諸問題とこれに関連する海洋学及び水産学上の諸問題について蘊蓄を傾け，また Lill 氏は特にチェサピーク湾の水産業とこれに関連する微細海洋学（Micro-Oceanography）上の諸問題について興味ある報告をなした。更に同日，丸の内の第一生命ホールで読売新聞社主催による両氏の一般講演会が開催され，Revelle 博士は“海底の神秘と海洋資源”と題し，該博なる知識を傾けた興味溢るる講演をなし，また Lill 氏は“米国の海洋学研究”と題し，合衆国における海洋学研究の概況を説明し，特に米国政府や公共団体等が海洋学研究にどのように力を注ぎ財政支出をしているかという点について詳述し，いずれも聴衆に多大の感銘を与えた。（中野猿人記）

編集後記 第 2，3 号には海洋循環に関連する論文が比較的多く収録されることになりました。これは，一つは，今秋 10 月東京で開催される予定のユネスコ国際海洋学シンポジウムの主題が“海洋循環”であるため，その方面の論文をこの 2 つの号に多く掲載しようという編集方針が取られたためです。なお，“*Records of Oceanographic Works in Japan*”（日本学術会議太平洋学術研究連絡委員会編集，日本ユネスコ国内委員会刊行）に寄稿された論文原稿のうち，海洋循環に関連の深いもの数篇は，同誌編集委員会の希望意向により著者の御承諾を得て本誌に掲載されることになつたことをお断りしておきます。 日本海洋学会誌編集委員会

Errata

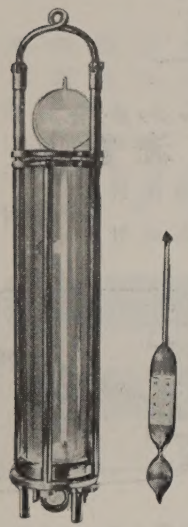
for Vol. 11, No. 2

N. WATANABE and T. HIRANO.

Page 51. In col. 1, line 16, for “**years**” read “**mouths**”.
In col. 1, line 21 from the bottom, for “**anormalies**” read “**anomalies**”.
In col. 1, line 13 from the bottom, for “**come**” read “**comes**”.

T.S.K. 鶴見精機

海洋調査観測用器機と航海計器



T. S トーメイ顚倒採水器
(寒暖計封入比重計付)

(捲上機) (測深儀) 航海用及海洋調査用。
(手　動) 250 M—550 M 三種アリ。
(携帯用) (河川、湖沼、石油井用) 200 M。T.S.O.O 型 ½ 馬力
T.S-B.T 用電動捲上機
(電　動) T.SO 型　550 M　2 馬力
T.S I 型　1500 M　2 馬力
T.S II 型　3000 M　3 馬力
T.S III 型　3000 M—5000 M　5 馬力及 10 馬力
T.S 海洋調査用ヘビータイプ電動捲上機 (15 馬力・30 馬力)
(自記験潮器) (水位計) 水圧式。フユース式。スチブンソン型。
(採水器) ナンセン式。エクマン式。北原B号。ゴム二重筒表面採水器
T.S トーメイ顚倒採水器 (プラスチック製南日、岩宮式)
T.S 精密採水器 (水平型)
(T.S 水深水温記録計) (T.S. バシサーモグラフ) 70。 150.300 M。
(顚倒寒暖計) 防圧、被圧、検定付
(油井用連装採水器) (蔵田式噴井用点灯回転計)
(水色計) T.S 水色計 (南日式)
(傾角度盤) 深度更正表付
(採泥器) 2 M 大型ピストンコーアサンプラー。ドレツヂ。
スナツパー式。岩宮式筒型。
(プランクトン採集器) 表層。中層用。植物プランクトン用。
元田、井上式垂直各層別採集器
中井、岩宮式曳航採集器 (濾水計、デプスレコーダー付)
中井式プランクトンネット用濾水計。
(深度計) 岩宮式2号、6 号。
(流速計) エクマンメルツ式。プライス電音式。
T.S 自記流向流速計。港研数点流速記録計。
T.S 隔測流向流速記録計。
南日式 遠隔電接流速計。
(波高計)
(測程儀) 特許 T.S 電気測程儀。舟尾測程儀。T.S サイドログ。
(旋回窓)
クリヤビユースクリーン)
○(其他の海洋調査観測機器)
○(T.S ステアリングスタンド。テレグラフ。其他の航海計器)

株式会社 鶴見精機工作所

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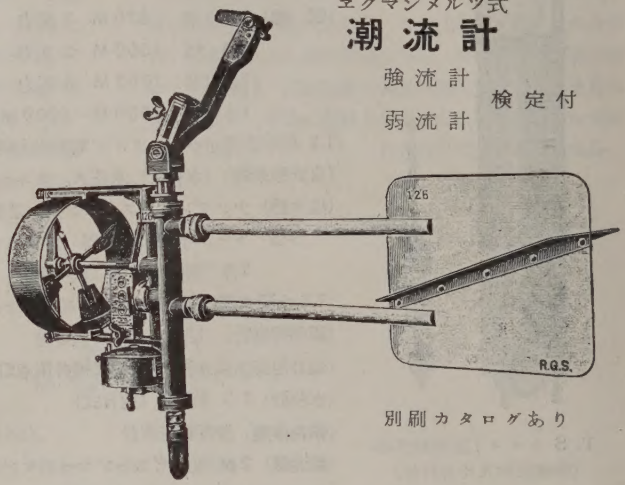
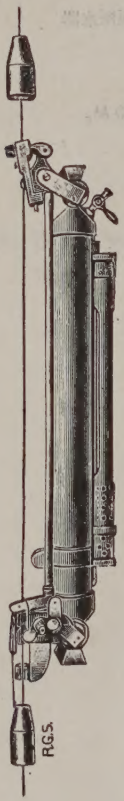
離合社の海洋觀測器械

エクマンメルツ式
潮流計
強流計
弱流計
検定付



別刷カタログあり



ナンゼン式連結転倒採水器

転倒寒暖計

營業品目

採水器	ナンゼン式（2種）エクマン式 特許北原式B号 特許KR式 特許神谷式航走用表面採水用
採泥器	新野式 丸川式 エクマンビルゲ湖沼用（3種） 田中式 新野式ドレツヂ（特許）
プランクトンネット	北原式 丸川式 電気中層用 中井式
測深機及關係測器	手動式 ルーカス式 気象台式 湖沼用 深度計（メートルゲージ） 傾斜測定器（表付）
潮流計	エクマンメルツ式 強流弱流検定付 河川用プライス流速計

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気象観測器械
海洋調査器械
顛倒温度計各種
（応需故障修理）
標準温度計各種
比重計各種

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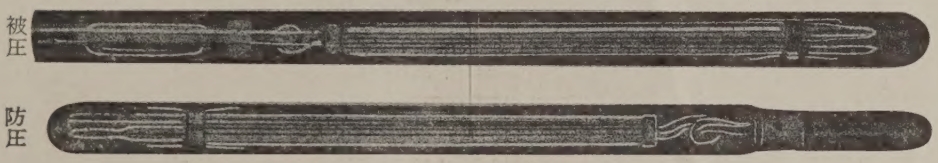

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